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Proceedings of the
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On the Replacement of Dolomite by Calcite in some Mesozoic Limestones from the French Jura

by D. J. SHEARMAN, J. KHOURI and the late S. TAHA

Read 3 July 1959

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ABSTRACT: Limestones are described in which dolomite is replaced by calcite. The petrographic evidence for the replacement is presented and the various textures are described.

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1. INTRODUCTION

THE REPLACEMENT of calcite by dolomite is a common feature in many limestones, but the subsequent replacement of the dolomite by calcite appears to be a relatively rarely described phenomenon. Von Morlot (1848) considered that the replacement of dolomite by calcite was a possible process in limestones and coined the term 'dedolomitisation'. J. J. H. Teall (1903) used the term to describe the metamorphic transformation of the mineral dolomite, and it has since become established in metamorphic petrology. The word is used in this paper, on the grounds of priority, however, in the manner originally proposed by Von Morlot.

Tatarskiy (1949) reviewed the occurrences of dedolomitised limestones described by Soviet geologists and records that they have been found in various parts of Central Asia and also in Lithuania. Cayeux (1935) discussed dedolomitisation as a possible process, but he did not give examples. Some of the rocks described by him (1916 and 1935) from the Senonian and Tertiary of the Paris Basin are closely similar to some of the dedolomitised limestones dealt with in this paper. Sander (1951) concluded from a study of limestone fabrics that extensive calcitisation of dolomite had occurred in the Northern Alps. Gilbert (1954) observed in general terms, but without referring to specific examples, that 'late calcite replacing dolomite is irregularly distributed through some carbonate rocks, the calcite embaying and transecting individual dolomite crystals'. Khvorova (1957 and 1958) has illustrated and given detailed petrographic descriptions of several varieties of limestones from the Carboniferous of the Russian platform in which dolomite has been replaced by calcite.

Fondeur, Grottis, Rouire & Vatan (1954) discussed the dolomitisation of the Jurassic rocks of France including the area dealt with in this paper. They recorded several phenomena which the present authors interpret as dedolomitisation, but on the balance of their evidence they rejected this possibility.

In the following paper the authors record the occurrence of rocks which they interpret as dedolomitised limestones from the Upper Jurassic and Lower Cretaceous of the French Jura in the southern part of the province of Ain and the adjacent parts of Savoie. Specimens were collected from widely scattered localities centred around Belley, extending from the Col du Chat and Conjux in the east, to Rossillon in the north-west and near Izieu in the south-west. The field relations of these rocks are under investigation and the purpose of the paper is to describe some of the petrographic types which have so far been found.

Favre & Richard (1927), in their studies of the Upper Jurassic of the Southern Jura, described some of the rock types occurring in the general area of La Balme but did not apparently record the occurrence of dedolomitised rocks.

2. THE NATURE OF THE EVIDENCE

Some 500 specimens of limestones and dolomites collected from the area have been examined in thin section; of these thin sections approximately 12 per cent show evidence of what the authors interpret as dedolomitisation. In one instance in the Kimmeridgian rocks of La Balme it was possible to follow, within one bed, the lateral change from dolomite into limestone after dolomite. In the majority of the other specimens the evidence for dedolomitisation rests essentially in the texture of the rock as seen in thin section, coupled with chemical analyses, and is shown in a variety of ways:

(i) By the occurrence of relics of incompletely replaced dolomite crystals.

(ii) By the presence of pseudomorphs of calcite after dolomite.

(iii) By the existence of palimpsest textures in which rhombic zones of ferric oxides or grain boundaries of the earlier dolomite crystals remain as ghosts within the new generation of calcite.

These lines of evidence are supported by the occurrence of rocks which are the dolomitic analogues of the dedolomitised ones.

3. TYPES OF DEDOLOMITISATION TEXTURE

It is convenient for the purposes of description to subdivide the dedolomitised limestones into three groups on the basis of the texture of the calcite replacement mosaic.

(1) Those rocks in which the individual crystals of dolomite have been replaced by a finer-grained mosaic of calcite crystals.

(2) Those rocks in which the dolomite has been replaced by a mosaic of calcite crystals which are coarser grained than the earlier dolomite.

(3) Those rocks in which the calcite replacement mosaic has tended to regenerate the pre-dolomitisation texture of the limestone.

Each of these types will be described in turn.

(a) Type 1. Dedolomitised Rocks in which the Individual Dolomite Crystals have been Replaced by a Finer-Grained Mosaic of Calcite Crystals

Several examples have been found of limestones carrying composite calcite rhombs (see Plate 1 A). These rhombs, the boundaries of which are sharply defined, do not consist of single crystals of calcite but are made up a mosaic of small crystals. In shape, these composite calcite rhombs differ in no way from dolomite rhombs, and the authors interpret them as pseudomorphs of calcite after dolomite. This interpretation rests on the observation that in dolomitised limestones the dolomite crystals are strongly crystalloblastic and, unless they mutually interfere, they invariably show good rhombohedral outlines; whereas calcite on the other hand does not appear to develop a rhombohedral habit in limestones. Cayeux (1916 and 1935) observed similar composite calcite rhombs in the Senonian of the Paris Basin, and concluded that the criterion based on rhombohedral habit was only in the nature of a general rule and that calcite can in fact occur as rhombohedra in limestones. The rocks from the Senonian were, he believed, examples of this. However, the composite calcite rhombs are not single crystals of calcite, and Cayeux justifiably rejected the possibility that the calcite had recrystallised on the grounds that when it does so it always produces a coarser-grained mosaic and never a finer-grained one. He suggested that the calcite rhombohedra had been removed by solution and the rhombohedral cavities subsequently filled with secondary calcite, i.e. that they are pseudomorphs of calcite after calcite. Such an explanation requires that the calcite rhombohedra be selectively dissolved relative to the calcite of the rest of the rock, which would be difficult to account for; and in any case the mosaics of the composite calcite rhombs described by Cayeux and also those in the Jura rocks do not show the characteristic features of the drusy mosaic described by Bathurst (1958).

Dolomite rhombs which contain cores of fine-grained calcite (Plate 1 B) are not uncommon in some of the fine-grained limestones. Considered on their own, these dolomite rhombs with calcite cores could have originated in two ways: either (i) by incomplete dolomitisation, in which case the calcite cores will be relics of the original limestone; or (ii) by partial dedolomitisation, in which case the calcite cores are a new generation of calcite which has replaced the dolomite. In the examples studied by the authors it was not possible satisfactorily to compare the texture of the calcite cores of the dolomite rhombs with the groundmass of the limestone; and although in the Jura rocks, for reasons which follow, the authors consider the phenomenon to be one of partial dedolomitisation they would hesitate to extend that interpretation beyond those rocks studied without other supporting evidence. Similar dolomite rhombs with calcite cores have been described by several authors, including Cayeux (1916 and 1935). Fondeur, Grottis, Rouire & Vatan (1954) observed them in the Jurassic of France and interpreted them as imperfect dolomitisation. Favre & Richard (1927) described dolomites from the Portlandian rocks of La Balme gorge in which dolomite rhombs contain inclusions of fine-grained calcite. Khvorova (1957 and 1958), on the other hand, interpreted the same phenomenon in the Carboniferous of the Russian platform as dedolomitisation.

In any one specimen the calcite cores are essentially of the same order of size, but it is possible to arrange the various specimens in an order (see Plates 1 B, 1 C and 1 A) which would give a series extending from normal dolomite rhombs on the one hand to composite calcite rhombs on the other. If this series is a true one it is difficult to read it in any other way than that it represents stages in the centrifugal replacement of dolomite by calcite.

Confirmation of the interpretation that the composite calcite rhombs are pseudomorphs of calcite after dolomite, and that calcite cores to dolomite rhombs represent partial dedolomitisation, was obtained from Kimmeridgian rocks of La Balme, where it was possible to trace the lateral change from dolomite into limestone after dolomite.

The Kimmeridgian rocks of La Balme gorge are massive limestones and dolomites which show a large-scale patchiness in the field resulting in part from the different colour, texture and weathered surfaces of the limestones

---

EXPLANATION OF PLATE 1

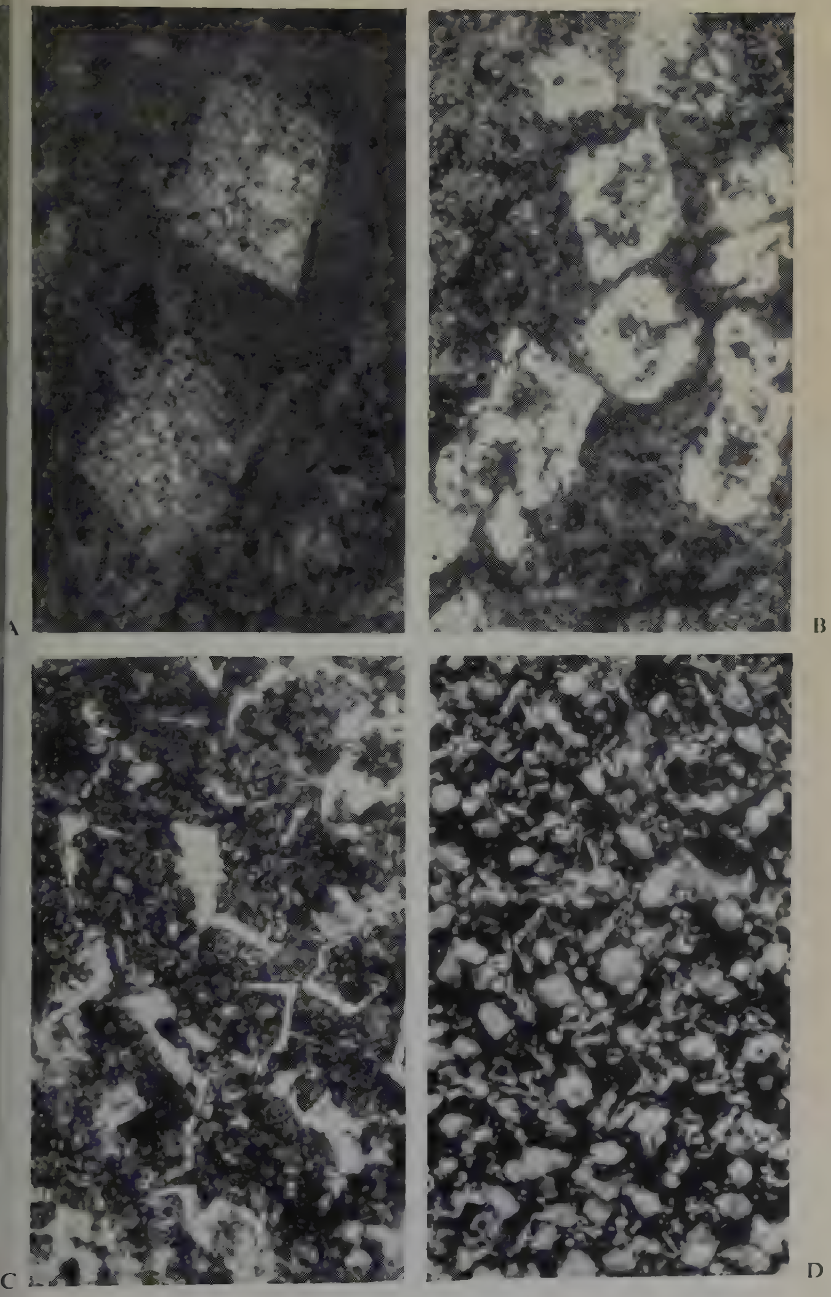
A. Composite calcite rhombs in fine-grained limestone

B. Dolomite rhombs with cores of fine-grained calcite. Dolomite colourless; calcite dark, stained with silver chromate

C. Relics of dolomite (colourless) with cores of fine-grained calcite (dark, stained with silver chromate)

D. Acid-etched polished surface of transition between dolomite (see Plate 2A) and limestone after dolomite (see Plate 2B). Kimmeridgian. La Balme gorge

PLATE 1



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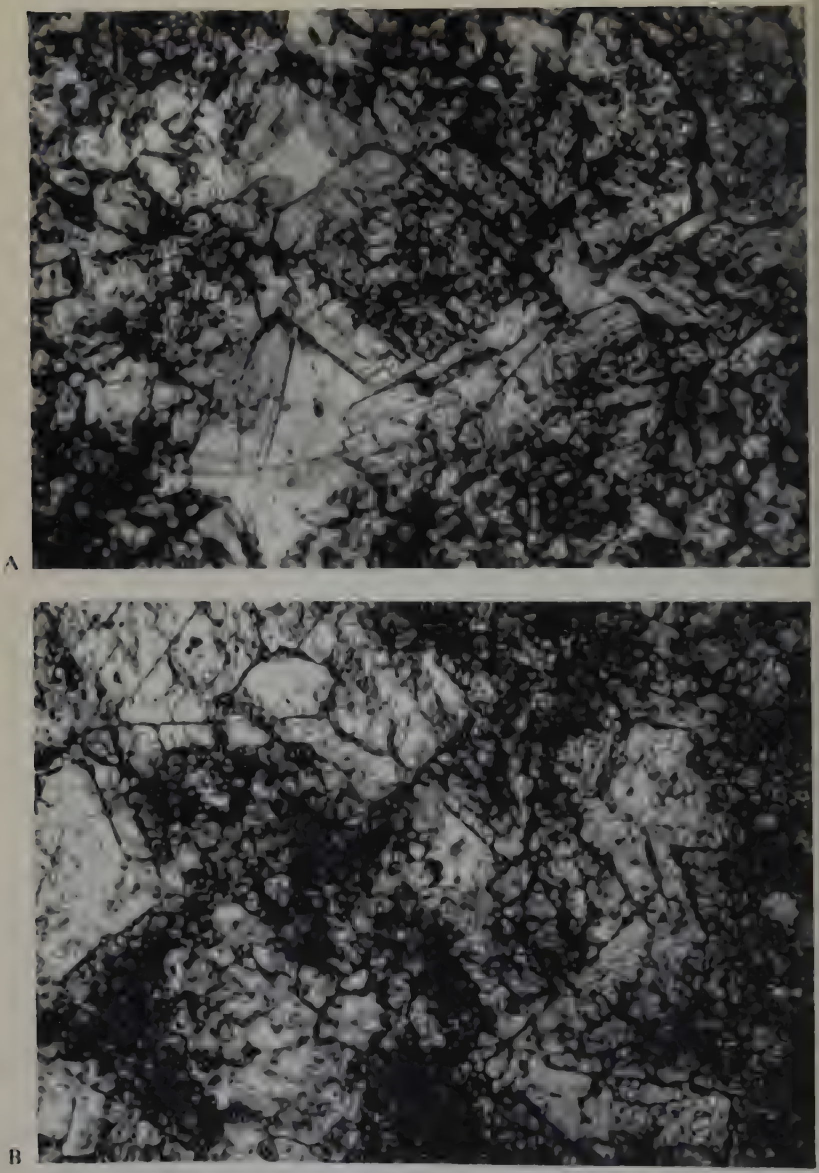

A Dolomite; Kimmeridgian, La Balme gorge

B. Limestone after dolomite; Kimmeridgian, La Balme gorge

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and dolomite. The transition between the areas of dolomite and limestone n any one bed is relatively sharp and is marked by a zone of ferric oxide staining one and a half inches or so wide. Three specimens were collected across one of the transitions from dolomite to limestone; one of the transition zone itself and one each from the limestone and dolomite on either side. Plates 2A and 2B are photo-micrographs of thin sections of the dolomite and limestone respectively. In texture the two rocks are superficially similar, both consisting of mosaics of mutually interfering rhombs of essentially the same order of grain size. In the dolomite rock (Plate 2A) the rhombs are rhombs of dolomite, with locally irregular infillings of calcite. In the limestone, on the other hand, the rock is entirely calcite (Plate 2B) and each of the rhombs is made up of a fine-grained mosaic of calcite crystals, and they are clearly pseudomorphs of calcite after dolomite. Across the transition from dolomite to limestone the dolomite crystals are progressively replaced by calcite and the complete change from one rock type to the other takes place within a distance of approximately one centimetre. Plate 1D shows an acid-etched polished surface of part of the transition zone, and it can be seen that the dolomite rhombs have developed rhomb-shaped cores and zones of calcite, apparently having been replaced centrifugally; and that these cores, in general, increase in size from the dolomite rock towards the limestone.

Plate 3A illustrates a similar limestone after dolomite from the Kimmeridgian near Chanaz.

Members of the Institut Français du Pétrole (Charpal de O., 1959) described a transition from limestone to dolomite in the Lusitanien of Cancanas, Hérault, which they interpreted as one of dolomitisation. The transition is marked in the field by a narrow zone of ferric oxide staining and in detail is characterised by dolomite crystals with cores and zones of calcite essentially similar to that described by the present authors from La Balme gorge. Thus it would appear that the transition zone between limestone and dolomite resulting from dolomitisation can be essentially similar to that between dolomite and limestone resulting from dedolomitisation. The evidence for dolomitisation versus dedolomitisation is found in the texture of the limestone. If this shows the texture of a normal sedimentary limestone then the transition is clearly one of dolomitisation; if, however, the limestone shows a relict texture after dolomite the transition is one of dedolomitisation.

This style of dedolomitisation has affected a variety of rock types ranging from those which were originally fine-grained limestones to oolitic limestones. In some of the fine-grained limestone where dolomitisation was patchy, with the dolomite crystals aggregated into a system of clots and stringers, dedolomitisation has produced a rock whose texture superficially resembles *structure grumeleuse*.

An interesting example of a dedolomitised oolitic limestone was found in the Kimmeridgian of the Swiss Jura at Ostscite. This rock shows dark fine-grained calcite ooliths set in a mosaic of coarser-grained calcite, which is seen to consist of an aggregate of composite calcite rhombs. The margins of the oolites are serrated, being penetrated by the rhombs.

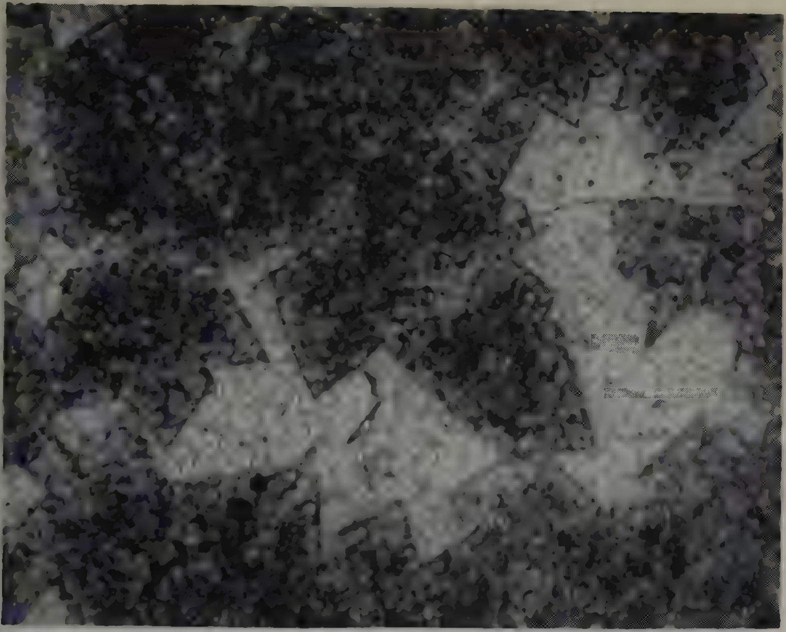
(b) Type 2. Dedolomitised Limestones in which the Dolomite has been Replaced by a Mosaic of Calcite Crystals which are Coarser Grained than the Earlier Dolomite

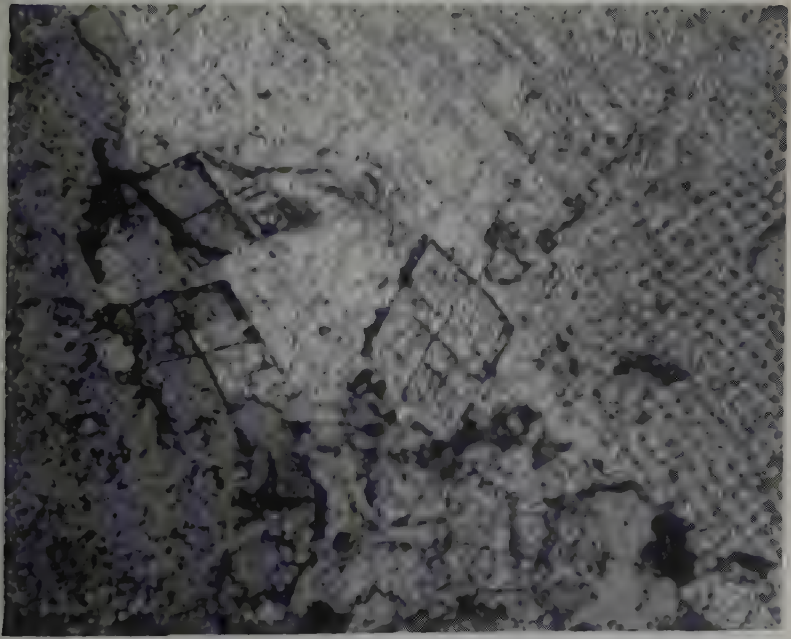
Locally within the succession beds of fine-grained granular dolomites are not an uncommon rock type, and in some of these there are narrow bands, essentially parallel to the bedding, which are much coarser grained and under the hand-lens show lustre mottling. Occasional thin irregular vein-like offshoots from these bands pass out into the finer-grained dolomite. In thin section these lustre-mottled rocks commonly consist of a relatively tightly packed aggregate of dolomite crystals poikilitically enclosed within larger crystals of calcite. Often the dolomite crystals are so tightly packed that the interspaces between are so small that even under crossed nicols the poikilitic texture of the interstitial calcite is not always immediately apparent. Fig. 1 illustrates the texture of one of these rocks. In this example the dolomite rhombs are less densely packed than is usual; the rhombs show clear sharp crystal boundaries, and have not been modified in any way. Commonly in these rocks the boundaries of dolomite crystals are irregular in outline and are embayed by the enclosing calcite crystals (see Figs. 2 and 3) and appear to have been partially replaced centripetally by them. Khvorova (1957 and 1958) described similar rocks in which the dolomite crystals are poikilitically enclosed by calcite, but in these rocks the dolomite rhombs contained cores of granular calcite. She regarded the rocks as partially dedolomitised. If this process of replacement by poikilitic calcite were to go to completion, the resulting limestone would be an essentially equigranular calcite rock. Such rocks do occur (see Fig. 4). In these the calcite crystals are optically clear and essentially free from inclusions, and their grain boundaries complex. In thin section these rocks have all the general features of metamorphosed limestones and they appear strangely out of place in a normal sedimentary sequence. Figs. 3 and 4 illustrate the textural affinity between these limestones and the poikilitic calcite of the dolomite rocks.

EXPLANATION OF PLATE 3

A. Limestone after dolomite. Kimmeridgian, Chanaz

B. Ferric oxide ghosts of dolomite rhombs in secondary calcite overgrowths on echinoderm fragments





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A variant of this type of dedolomitisation texture is illustrated in Figs. 5 and 6. Thin sections of these rocks seen under ordinary light (Fig. 5) show a granular mosaic in which straight-grain contacts and occasional acute and obtuse angles suggest that this is a dolomite mosaic. The grain boundaries are not sharply defined, but somewhat diffuse, and are outlined by ferric oxide staining. The rock, however, consists entirely of calcite, and under crossed nicols it is seen to be composed of a coarser-grained mosaic of calcite crystals (see Fig. 6), each of which encloses a number of the ghost dolomite crystals. The authors interpret these limestones as dedolomitised rocks in which the outlines of the earlier dolomite grain boundaries are preserved as a palimpsest texture.

(c) Type 3. Dedolomitised Limestones in which the Calcite Replacement Mosaic has Tended to Regenerate the Pre-dolomitisation Texture of the Limestone

A number of specimens have been found which show this type of replacement texture. They are all bioclastic calcarenites. Cementation of these rocks is by secondary overgrowth of calcite against monocrystalline echinoderm debris, and by the normal type of calcite pore-filling mosaic against the polycrystalline shell debris. Within the calcite of the cement there are rhombic zones of ferric oxides which in their outline closely resemble the zones of ferric oxides which often occur within dolomite rhombs. These ferric oxide zones are interpreted as the ghosts of former dolomite crystals which have been replaced by calcite. Where these zones of ferric oxide occur within the calcite overgrowths on echinoderm debris (see Plate 3B) the calcite is crystallographically continuous across the whole area, outside and within the rhomb. Locally the rhombic zones lie athwart the boundary between two secondary overgrowths, which passes unbroken across them. A thin section of this rock was treated with 1 per cent hydrochloric acid and potassium ferricyanide to test for the presence of iron-rich and iron-poor calcite. The calcite of any one fragment of echinoderm debris, the overgrowth on that fragment and the included rhombs, although all one single crystal of calcite, stained differentially. The calcite of the echinoderm fragment took on a faint irregular blue stain; the overgrowth stained strongly blue; but the calcite enclosed within the rhombic zones of ferric oxide remained unstained. Further, outside the ferric oxide zones there remained a narrow rhombic zone of unstained calcite. This differentiation of iron-rich and iron-poor calcite helps to elucidate the diagenitic history of the rock. The first phase was the development of relatively iron-rich calcite overgrowths on, and in lattice continuity with, the calcites of the detrital echinoderm fragments. The paler, patchy distribution of the stain on the echinoderm fragment may result from the fact that although the organic fragment was itself relatively iron free, the canals which may

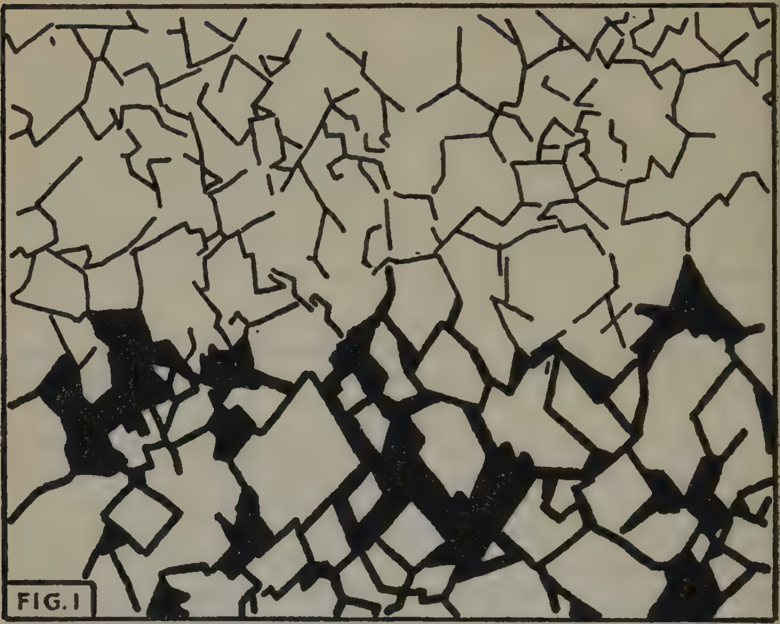

FIG. 1

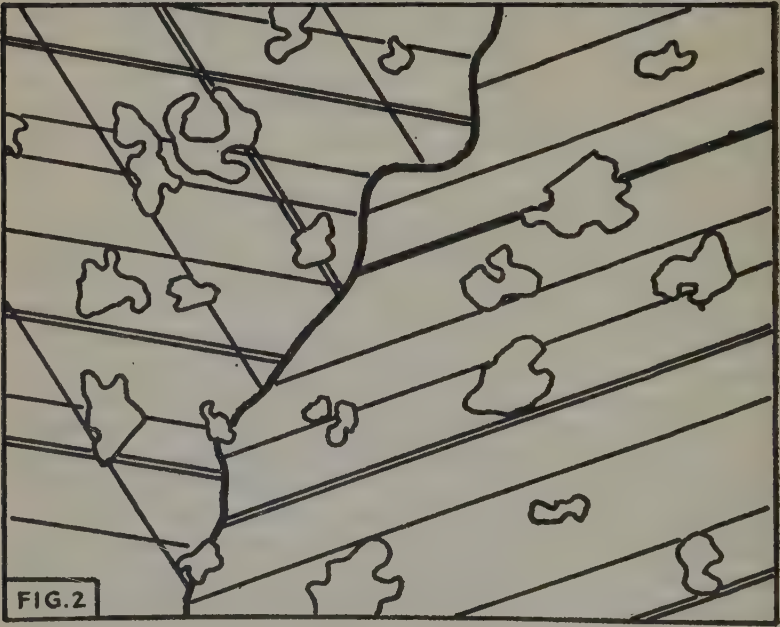

FIG. 2

make up 50 per cent or so of its volume are filled with iron-rich calcite deposited at the same time as the overgrowth. The rock was subsequently dolomitised, and scattered dolomite rhombs locally replaced the iron-rich calcite of the overgrowths. Later, the process of dedolomitisation replaced the dolomite by an essentially iron-free calcite. The limits of the earlier dolomite rhombs are marked by the boundaries between the stained and unstained calcite, and the rhombic zones of ferric oxides represent zones within the dolomite.

Where the rhombic zones of ferric oxides occur within the calcite of the pore-filling mosaic, the calcite crystals pass uninterrupted across them, and the texture of the mosaic is essentially similar to that in those pore spaces which do not carry the dolomite palimpsests. Thus it would appear that the dolomite replaced the pore-filling mosaic and that dedolomitisation has tended to regenerate that earlier mosaic.

4. CONCLUSIONS

Where partial dedolomitisation of a rock has occurred and unreplaced relics of the dolomite crystals remain, the evidence from the Jura rocks suggest that the replacement process may be either centrifugal or centripetal. These conclusions are in agreement with those of Khvorova (1957, 1958). Where centrifugal replacement has taken place the individual dolomite crystals are replaced by a mosaic of finer-grained calcite, i.e. the dolomite is pseudomorphed by calcite. This type of replacement has affected a variety of rocks ranging from dolomitised limestone in which the dolomite rhombs are scattered throughout the limestone groundmass, to those which were entirely a mosaic of dolomite crystals. In the rocks where centripetal replacement can be demonstrated to have taken place the replacement was by larger calcite crystals which poikilitically enclosed the dolomite. This type of replacement has only affected rocks which were essentially dolomite and which consisted of a tightly packed aggregate of dolomite crystals.

Khvorova described rocks where the dolomite crystals are poikilitically enclosed in calcite and in which the dolomite rhombs have cores of fine-

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EXPLANATION OF FIGURES 1 AND 2

Fig. 1. Dolomite crystals poikilitically enclosed within larger crystals of calcite. The black part of the field represents a large calcite crystal in the extinction position (high power magnification). Portlandian–Purbeckian; Rossillon, Ain

Fig. 2. Dolomite crystals partially replaced by poikilitic calcite (high power magnification). Portlandian–Purbeckian; Rossillon, Ain

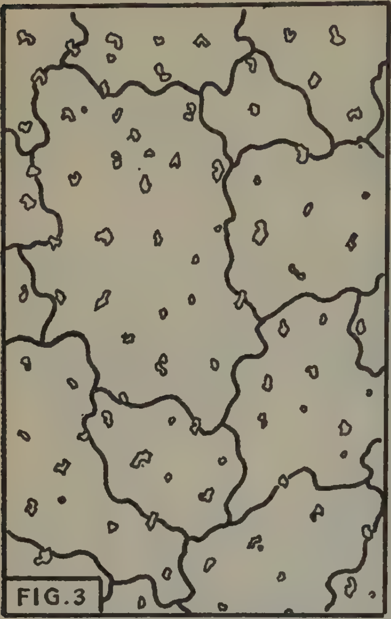

FIG. 3

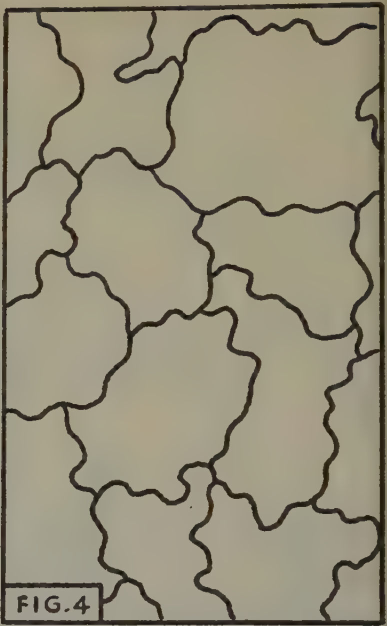

FIG. 4

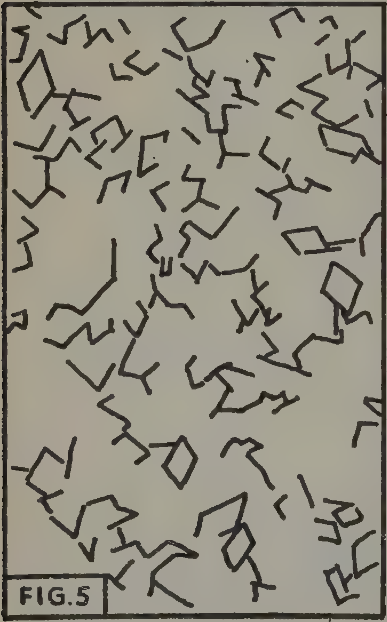

FIG. 5

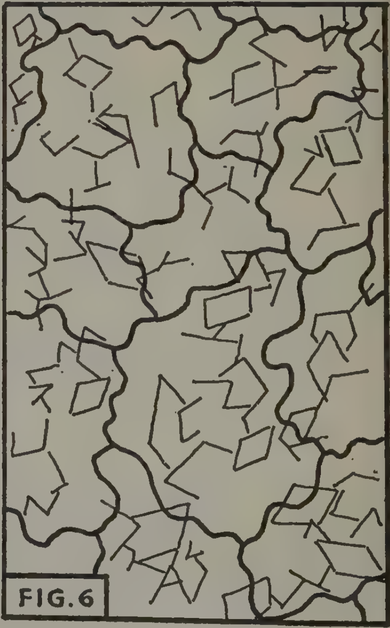

FIG. 6

grained calcite. No examples of this type have so far been found by the present authors.

Dedolomitised and partially dedolomitised rocks have been found intimately associated in the field with dolomitic rocks which are unaffected. In some of the rocks dedolomitisation has gone to completion, in others it is only partial. The phenomenon is widespread in the area under discussion, and the occurrence of similar rocks in the Swiss Jura suggests that dedolomitisation may have been developed on a regional scale.

Von Morlot (1848), discussing the possible role of magnesium sulphate solutions in the formation of dolomite, according to the equation:

$$2\ \mathrm{CaCo_3} + \mathrm{MgSo_4} = \mathrm{CaCo_3\ MgCo_3} + \mathrm{CaSo_4}$$

observed that this is a reversible reaction, and suggested that gypsiferous solution could bring about dedolomitisation.

Tatarskiy (1949), discussing the Russian occurrences of dedolomitised limestones, stated that it is always a near surface phenomenon and that borehole samples do not show any evidence of dedolomitisation at depth. Khvorova (1958) attributed the dedolomitisation of the Carboniferous rocks of the Russian platform to the action of salts washed out from the Lower Permian rocks. Tatarskiy recorded an occurrence of dedolomitised limestones where the rocks show an efflorescence of epsomite. This is a seasonal mineral and is washed away during the rainy season, and Tatarskiy concluded that in this region the process of dedolomitisation is operating at the present day.

The reaction suggested by Von Morlot is a very likely one. In the Jura region superficial gypsiferous solutions could have been available; and there is also the possibility that suitable solutions could have been mobilised from the underlying evaporites. However, until the field relations of these rocks are better known and subsurface data available, the speculation on the causes of the dedolomitisation of the Jura rocks is best left in abeyance.

EXPLANATION OF FIGURES 3 TO 6

Fig. 3. Relics of dolomite crystals poikilitically enclosed in calcite (low power magnification). Portlandian–Purbeckian; Rossillon, Ain

Fig. 4. Limestone after dolomite (low power magnification). Portlandian–Purbeckian; Rossillon, Ain

Fig. 5. Limestone showing ferric oxide ghosts of earlier dolomite rhombs (low power magnification). See Fig. 6. Portlandian–Purbeckian; Rossillon, Ain

Fig. 6. The same specimen as Fig. 5 but showing calcite grain boundaries (low power magnification)

The late Said Taha, one of the authors of this paper, died a month before the paper was read to the Association. His home was near Damascus in Syria. He had been a student in London and died during the final year of his post-graduate research. Douglas Shearman and John Khouri hope that this paper would serve as a memorial to their friend Said Taha.

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DISCUSSION

PROFESSOR J. H. TAYLOR congratulated the authors on their interesting paper and on the excellence of their photo-micrographs. Dedolomitisation had been noted by Teall as a metamorphic effect in the Durness Limestone near the Cnoc na Croine intrusion and it was not uncommon in wall rock alteration along hydrothermal veins. It was only relatively recently, however, that its importance as a low-temperature process was appreciated. It had not so far been recorded from any British dolomites but the speaker had little doubt that it would be recognised now that the replacement textures involved had been so clearly described by Mr. Shearman and his colleagues. Somewhat similar textures occurred where calcite replaced siderite in sedimentary ironstones. This replacement had generally been regarded as an ancient diagenetic change but the possibility of its being in part a superficial effect must now be borne in mind.

Joint-Drags in Mid-Argyllshire

by J. L. KNILL

Received 11 March 1960

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ABSTRACT: Joint-drags are angular fold-structures, with steep axial planes, which frequently appear to have been developed during faulting. In Mid-Argyll, the fractures which form the axial planes to the joint-drags trend consistently NW.–SE. and dip steeply south-westwards; there is a dextral sense of movement along these fractures. It is suggested that the joint-drags were formed as a series of slip structures complementary to the sinistral NNE.–SSW. tear-fault system of the SW. Highlands of Scotland.

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1. INTRODUCTION

THE OCCURRENCE of folds with an angular profile, which appear to be associated with movements on adjacent fault-surfaces, is a feature of deformed rocks. Flinn (1952) has described structures of this type from Unst which he terms joint-drags. A recent survey of large area of Dalradian rocks in Mid-Argyll has demonstrated that the orientation and style of these structures show a remarkable consistency. The present paper is concerned with the description of the form and origin of the joint-drags in this district.

2. DESCRIPTION OF THE JOINT-DRAGS

Joint-drags are best developed in thinly foliated rocks and, particularly, in slates and phyllites; however, they also occur in the more massive quartzites, grits and metamorphosed igneous rocks. In general, a joint-drag is made up of two steeply inclined planar fractures cutting across and offsetting the foliation (Fig. 1 a). The material caught between these fractures is disoriented so that the fold-shape is markedly angular. The intersection between the fractures and the foliation results in the development of a linear structure whose plunge is controlled by the attitude of the intersecting S-surfaces. Adjacent joint-drags, developed on differently inclined foliation surfaces, plunge at entirely different angles; thus, the attitude of these folds is not of any significance in a tectonic analysis of joint-drags. Therefore, the most important features of joint-drags are the attitude of, and sense of movement along, the fracture-surfaces.

The relative development of joint-drags varies in intensity from minute crenulations on cleavage faces to large corrugations associated with small-scale folds. Similarly, there are variations in the distribution of these structures and it is apparent that joint-drags tend to occur in NNE.–SSW. zones running parallel to the strike of the country rocks. The dimensions of

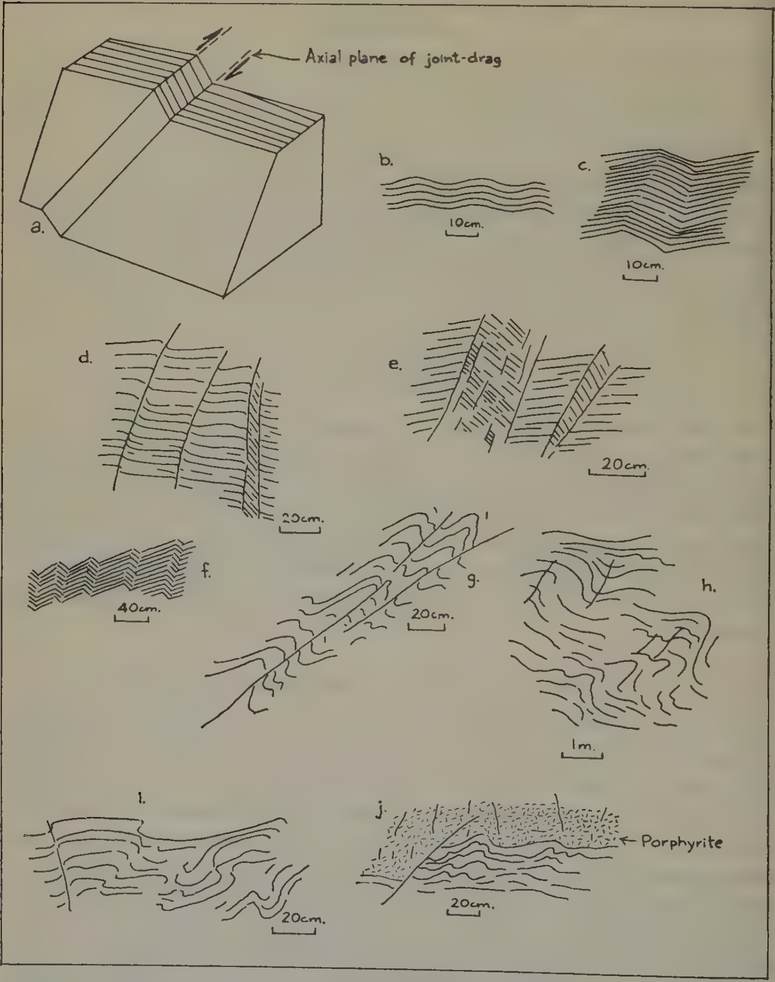

Fig. 1. (a) Block diagram illustrating the general form of a joint-drag. (b) to (e) illustrate the sequence in development of joint-drags with the successive folding and fracturing of the fold-limbs. (f) Typical series of joint-drags. (g) Thrusts associated with joint-drags. (h) Large folds related to joint-drags. (i) Conjugate joint-drag. (j) Folding in base of porphyrite sill next to joint-drags. (b) to (j) are plan views of outcrops

these zones vary considerably and they can often be traced for a kilometre or more along the strike and may be up to 100 metres in width. The joint-drags are generally well developed in the centres of the zones and there is a gradual diminution in their abundance outwards. Occasionally, the joint-drags appear to be associated with small NNE.–SSW. faults. The joint-drags are sometimes developed as a faint crinkling on cleavage surfaces in the more micaceous phyllites. These crinkles are of the order of 0.2 mm. across; they vary in distance apart and five or more to the centimetre are frequently present. When examined in thin-section, it is seen that these microscopic joint-drags are similar to strain-slip cleavages; the micaceous minerals are strongly curved within the drag, but not broken. However, in slightly bigger structures the flexuring of the foliation is associated with ruptural fractures; occasionally, these slightly larger structures cut across the minute joint-drags and are, thus, formed at a later period.

On a still larger scale, joint-drags appear as a prominent symmetrical warping of the foliation. The wavelength of these structures varies between thirty and ten centimetres, with an amplitude of five to eight centimetres; locally, larger warps, up to twenty metres in wavelength, are present. Small-scale asymmetrical warps appear to develop readily from the symmetrical structures. When the asymmetry of the fold becomes marked, fractures generally occur on the shorter fold-limb resulting in the formation of a typical joint-drag. This sequence of events in the development of a joint-drag, based upon field sketches, is shown in Fig. 1 b–e. The relative distribution of joint-drags varies considerably, and in some areas they may occur at intervals of thirty to sixty centimetres along the strike (Fig. 1 f).

In extreme examples, movement continued after the formation of the joint-drag and a series of angular folds have resulted from the thrusting of one side of the structure past the other (Fig. 1 g). Occasionally folds up to five metres across have developed during these same movements (Fig. 1 h); joint-drags occurring within these structures are distorted and the folds, therefore, appear to be late in the deformation history.

About 1500 measurements of the attitude of the steep fractures, associated with the joint-drags, have been made in Mid-Argyll. These data ndicate that the fractures bounding the joint-drags strike consistently NW.–SE. and dip steeply towards the south-west; observations illustrating the trend of these fractures are shown in Fig. 2 a. A stereographic diagram, based on 104 measurements in an area of about four square kilometres, is also shown in the same diagram (Fig. 2 b); most of the fractures dip at between 76° and 87° SW. There are some local variations in the attitude of these structures and, occasionally, the fractures may dip at an angle as low as 50° to 60°. The sense of movement along these fractures (Fig. 1 a) is consistently dextral, that is, the north-easterly side moved southward relative to the south-westerly side (Fig. 2 a). Very occasionally, sinistral

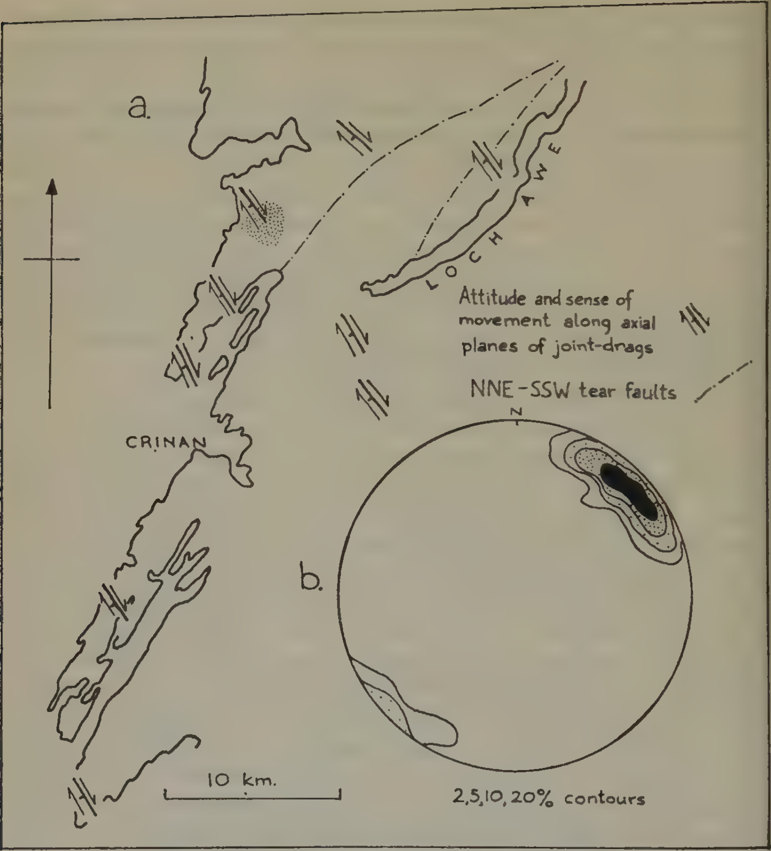
a.
LOCH AWE
Attitude and sense of movement along axial planes of joint-drags
NNE-SSW tear faults
CRINAN
b.
N
10 km.
2,5,10,20% contours

Fig. 2. (a) Map of Mid-Argyll showing the orientation and sense of movement along the axial planes of joint-drags. (b) Stereographic projection based upon 104 measurements of the axial planes of joint-drags from the area stippled in map (a) above

joint-drags are present either in isolated zones or combined with the usual type of joint-drag to form conjugate structures (Fig. 1 i).

3. ORIGIN OF THE JOINT-DRAGS

The Etive porphyrite dyke swarm, of Old Red Sandstone age, trends across Mid-Argyll and was emplaced prior to the formation of the joint-drags. Individual members of this swarm are cut by the steeply inclined fractures and, occasionally, the dykes are slightly contorted by the joint-drags (Fig. 1 j); thus, the joint-drags were formed after the extrusion of the Lorne Lavas and the emplacement of the Starav Granite (Anderson, 1937).

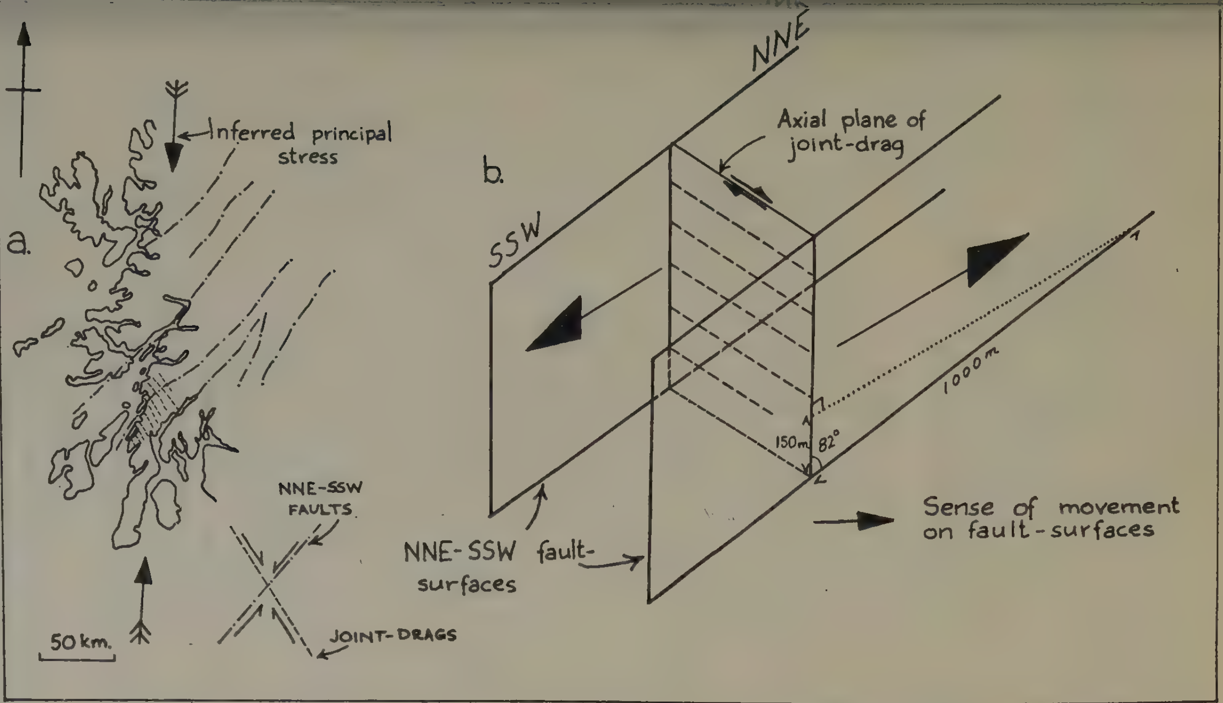
a.
Inferred principal stress
NNE-SSW FAULTS
JOINT-DRAGS
50 km.
b.
NNE
SSW
Axial plane of joint-drag
1000 m
150 m
82°
NNE-SSW fault-surfaces
Sense of movement on fault-surfaces

Fig. 3. (a) Map of western Scotland illustrating the relationship between the NNE.–SSW. tear-fault system and the axial planes of joint-drags. The principal stress system (oriented approximately north–south) is taken from Kennedy (1946). (b) Diagram showing the relation between the dip of the axial planes of the joint-drags and the NNE.–SSW. faults. An average dip of 82° for these axial planes would suggest a downthrow to the ESE., associated with a vertical displacement of 150 m. for every 1000 m. of horizontal movement

The joint-drags are cut by NW.–SE. faults and the Tertiary dykes. Many of the NW.–SE. faults of Mid-Argyll are later than the NNE.–SSW. fault system of the Scottish Highlands; there is little evidence as to the absolute age of the NW.–SE. faults, but a fracture with a similar trend displaces the Upper Carboniferous in southern Kintyre. The NNE.–SSW. faults displace both the Lorne Lavas and the porphyrite dyke swarm; the main movement on these faults is possibly Upper Old Red Sandstone or Lower Carboniferous in age (Kennedy, 1946). This evidence, together with the occurrence of joint-drags along small NNE.–SSW. faults, suggests that the joint-drags were developed at much the same period as the main NNE.–SSW. fracture system of the Highlands.

However, the relationship between the joint-drags and the NNE.–SSW. faults becomes clearer when the stress systems under which they were developed are considered. Several of these tear-fractures, including the Great Glen, Ericht–Laidon–Craignish and Tyndrum faults, cut across the SW. Highlands and are frequently associated with sinistral displacements (Kennedy, 1946). It has been suggested by Kennedy (1946) that the regional stress system which resulted in this fracture system was oriented approximately north–south (Fig. 3 a). In a system of this type, one might expect to find a complementary set of NW.–SE. dextral fractures; it appears likely, from the form and age relationships of the joint-drags, that these structures represent this complementary set (Fig. 3 a). Therefore, the dextral joint-drags of Mid-Argyll appear to be the direct result of sinistral movements along the tear-fault system of the SW. Highlands.

The consistent south-westerly inclination of the fractures, developed with the joint-drags, suggests that the tear component of the NNE.–SSW. faults was associated with a downthrow to the south (Fig. 3 b). Assuming an average dip of 82° for these steep fractures, this downthrow would be of the order of 150 m. for every 1000 m. of horizontal displacement. It is of interest to note that Leedal & Walker (1955) have demonstrated a similar relationship between the vertical and horizontal displacements (2000 foot downthrow with 2.2 miles lateral movement) on a NE.–SW. tear-fault, of an equivalent age, at Barnesmore, Co. Donegal. It is apparent, from observations by the author, in both Dalradian and Lower Palaeozoic rocks in other parts of the British Isles, that steep, south-westerly dipping fractures, forming the axial planes to dextral joint-drags, occur elsewhere. This tends to suggest that a stress system, similar to that represented in Mid-Argyll, may have been developed in other areas.

ACKNOWLEDGMENTS

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Denudation Chronology of the South-West Pennine Upland

by R. H. JOHNSON and R. J. RICE

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ABSTRACT: The western flank of the Pennines derives unity and individuality from the rapidly alternating grit and shale outcrops which produce highly characteristic landforms. Although the region is drained by several rivers, it is shown that there is a widespread distribution of erosion surfaces which can be grouped into a chronological sequence which is common to all. The drainage pattern of the area is critically examined and its significance with reference to the current hypotheses of superimposition from a Cretaceous cover or from high level marine surfaces during a succession of falls in sea-level is discussed.

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1. INTRODUCTION

IN A number of papers since 1950 Professor D. L. Linton has consistently claimed that the river pattern of much of Britain is inherited from an initial development on an emergent Cretaceous rock cover which was tilted eastwards. Such an assertion has not been left unchallenged and indeed has excited considerable controversy amongst British geomorphologists. In particular his views have been challenged by J. B. Sissons who has maintained that in Yorkshire the present drainage system developed from mid-Tertiary times by the extension of consequent streams across marine platforms created by an intermittent but continual fall in sea-level.

The studies by the above authors have been concerned primarily with east-flowing streams. Both accept that many of these streams are superimposed but differ as to the nature of the cover from which superimposition took place. In seeking to resolve this conflict much attention has been paid to the form and nature of the erosion surfaces on the interfluves between the east-flowing streams of south-west Yorkshire and north Derbyshire. Examination of these surfaces has not proved decisive and it is suggested that work might now be concentrated in areas where the two hypotheses are in sharp conflict and yet which are within reasonable proximity to the areas already considered. Clearly one such area will be the western flank of the south Pennines for this lies to the west of the main water-parting of Britain and is drained wholly by west-flowing streams. If Sissons' views are correct the conditions in this area should be comparable to those which he has described from south-west Yorkshire. Indeed, in his diagrams he has claimed the presence of an old shoreline on the western side of the Peak, Bleaklow and Black Hill in support of his arguments. It is not so easy to anticipate the conditions likely to be present if Linton's arguments prevail but in either case it is necessary to examine the evidence itself before further conjectures can be profitably considered.

The present paper considers the evidences for an area of some 190 square miles drained by the upper courses of four Cheshire streams—the Goyt, Dean, Bollin and Dane (Fig. 1). The method employed is, firstly, to describe the erosion surfaces possessing the attributes of marine platforms, and, secondly, to examine the relationship between structure and drainage for any light it may shed on the early phases of denudation.

2. THE EROSION SURFACES OF THE SOUTH-WESTERN PENNINES

The area drained by the Goyt, Dean, Bollin and Dane is composed almost entirely of the Millstone Grit and Coal Measure Series (Fig. 2). Both these formations consist of a great thickness of alternating grits, sandstones, shales and, especially towards the top of the Series, coal seams. The great contrasts between the various beds in their resistance to erosion has resulted in the geological structure being very clearly reflected in the landscape. Nevertheless, the landforms retain in their detailed configuration evidence of a long erosional history. In order to decipher the evidence it is necessary to distinguish between those features owing their form primarily to geological structure and those features indicative of significant phases in the geomorphological evolution. An erosion surface when first formed will pass from one outcrop to another with little change in form or gradient. But such a surface, once subjected to modification by weathering, will rapidly lose its simplicity of form and may often be completely destroyed where cut across the less resistant rocks. As a consequence it is to be expected that in the south-western Pennines evidence for an intermittently falling base-level—whether it be marine platforms or valley benches—will normally be present on the gritstone outcrops. As the gritstone is also the rock most likely to form structural benches the effects of the structure are difficult to disentangle from those of erosion and recourse must often be had to the overall distribution rather than to the detailed form of each

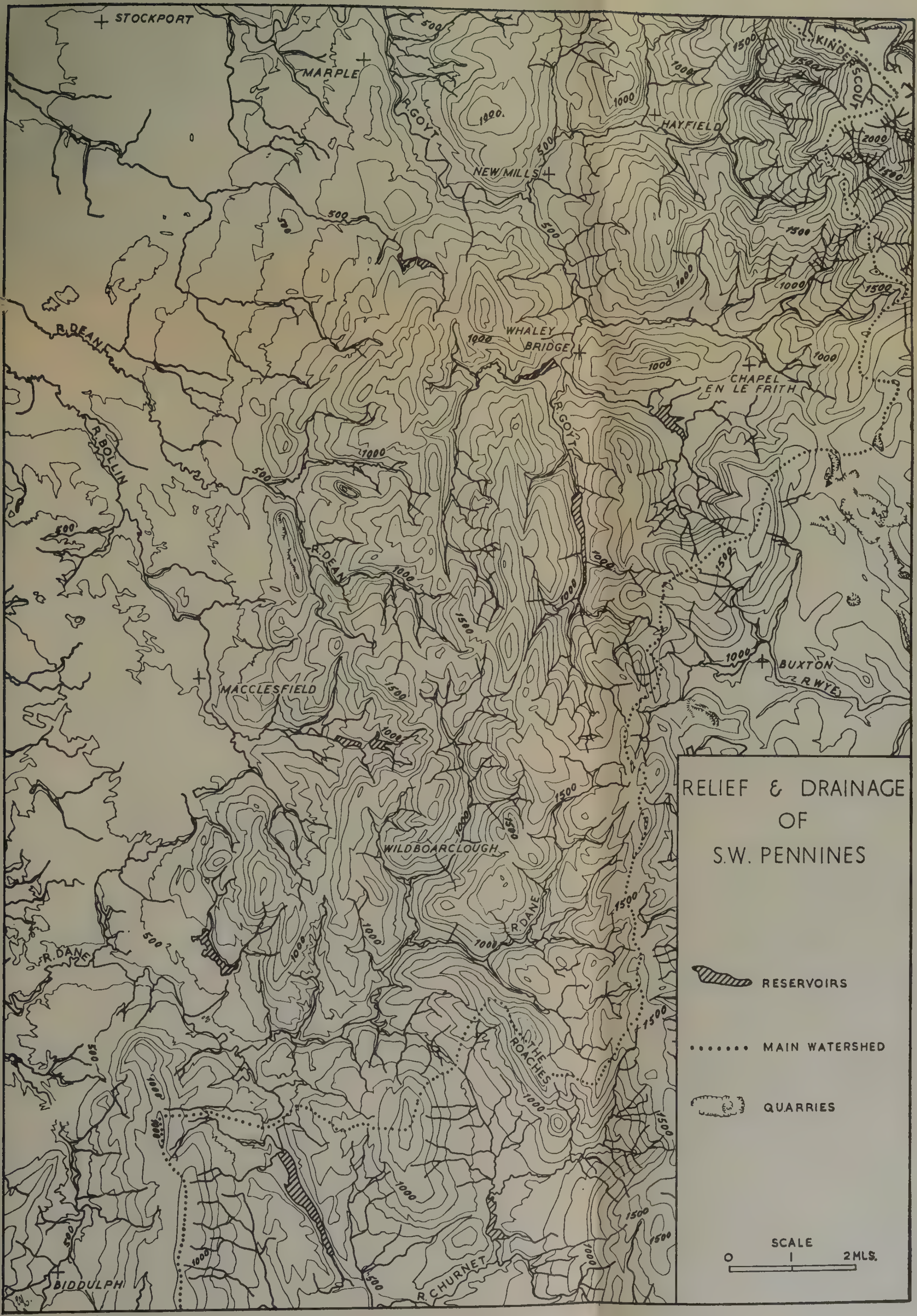

Fig. 1. Relief and Drainage of SW. Pennines

[*To face p.* 22

individual example. In view of this fact it was considered satisfactory to delimit the 'flats', and to estimate their heights, from information contained on the O.S. 1 : 25,000 maps; to map at the six-inch scale and to establish heights by aneroid survey seemed unnecessary. The degree of error possible in this method might be considered rather large for geomorphological investigations of this kind. Accurately surveyed contours are drawn at 100-foot intervals so that the theoretical error of interpolated contours could be as great as seventy-five feet. The method certainly would be unsatisfactory if employed in lowland areas for the mapping of river terraces or associated features. In this upland area the valley benches and associated upland features are widely spaced in their altitudinal distribution and thus there seemed little danger in using the sheets of the map available at this scale. Further, rough checks were obtained by reference to spot heights, etc., and also an ex-R.A.F. altimeter gave height measurements to within twenty feet.

(a) The Upland Plain

Without doubt the most striking erosion feature in the south-western Pennines is that which we term the Upland Plain. The existence of this surface was briefly noted by Sissons (1954, 325) and was tentatively correlated with his Bradfield Surface. From any vantage point above 1200 feet the widespread extent of the Upland Plain is immediately apparent and detailed mapping has shown that, in one area or another, it is developed on all the major grit bands of the Millstone Grit and Coal Measure outcrops. It is difficult to demarcate exactly its upper and lower limits for the surface is undulating rather than flat. In general it appears to slope from a height of about 1400 feet in the east, where it abuts residual masses of higher ground, to about 1200 feet at its most westerly points. Although best preserved east of Bollington and south-west of Shutlingsloe it is probably also represented by the accordant summits of Bosley Minn, Croker Hill and Gun Hill. On the surface itself the major grit bands often cause gentle swellings which may rise 100 feet or more above the surrounding remnants. As a result the dividing line between the irregularities of the surface and the higher residual masses becomes blurred. There appear to be four main residual masses, centred on Kinderscout, Axe Edge, Shutlingsloe and Macclesfield Forest.

By suggesting a correlation with the Bradfield Surface, Sissons implies that the Upland Plain is a marine platform. However, in the view of the present writers, the configuration of the Upland Plain accords more with its being the product of sub-aerial erosion. Admittedly there is a break of slope and fairly steep gradient at the inner margin of the surface (criteria frequently used in the recognition of marine platforms and an argument reiterated by Sissons (1954, 22)), but on the other hand the break of slope is not at a constant height, a fact which seems to the present writers of even greater importance. Moreover, the surface fingers into the residual masses along the valleys, a tendency well exemplified in the upper Goyt Valley. Further evidence on the origin of the Upland Plain is furnished by the drainage pattern, but consideration of this is deferred until the other erosion surfaces of the south-western Pennines have been described.

(b) The Valley Benches

At the western extremity of the Upland Plain the ground slopes steeply to the level of the Cheshire lowland so that below 1200 feet there is not one of the prominent fringing shelves which are so distinctive a feature of the Pennine landforms of south-west Yorkshire. Almost all the evidence for the later stages in the history of the region is contained in the valley benches which occur in the deep, steep-sided valleys incised into the Upland Plain; these stages may be briefly described for the light which they shed, albeit indirectly, on possible early shorelines.

The sequence of benches in the four river basins is remarkably alike, and reveals that rapid downcutting by the rivers was punctuated by at least three phases of stability and valley widening (Fig. 3). The first stable phase occurred after the rivers had incised themselves several hundred feet into the Upland Plain; in the Goyt and Dane Valleys there is evidence of interruption of this phase by renewed downcutting to a depth of fifty feet. The second period of stability occurred when the rivers had lowered their courses a further 150 feet, and the third when they had lowered them yet another 100 feet. The similarity of the physiographic records of the four river basins is immediately apparent from Table I, in which the estimated heights are given for the trunk rivers where they leave the high ground:

TABLE I

	Estimated river height (in feet O.D.) during the still-stands which succeeded incision into the Upland Plain		
	First stillstand	Second stillstand	Third stillstand
Goyt at Marple Bridge	(Dual) 940 and 890	690	580
Dean at Bollington	900	760	660
Bollin at Sutton	900	750	650
Dane at Bearda	(Dual) 940 and 890	750	obscured by drift

For the purposes of the present paper it will be sufficient, instead of giving a detailed description of each valley, to supplement the information contained in the above table, and in Fig. 3, by a few general observations on the valley benches. From its source to Whaley Bridge the River Goyt

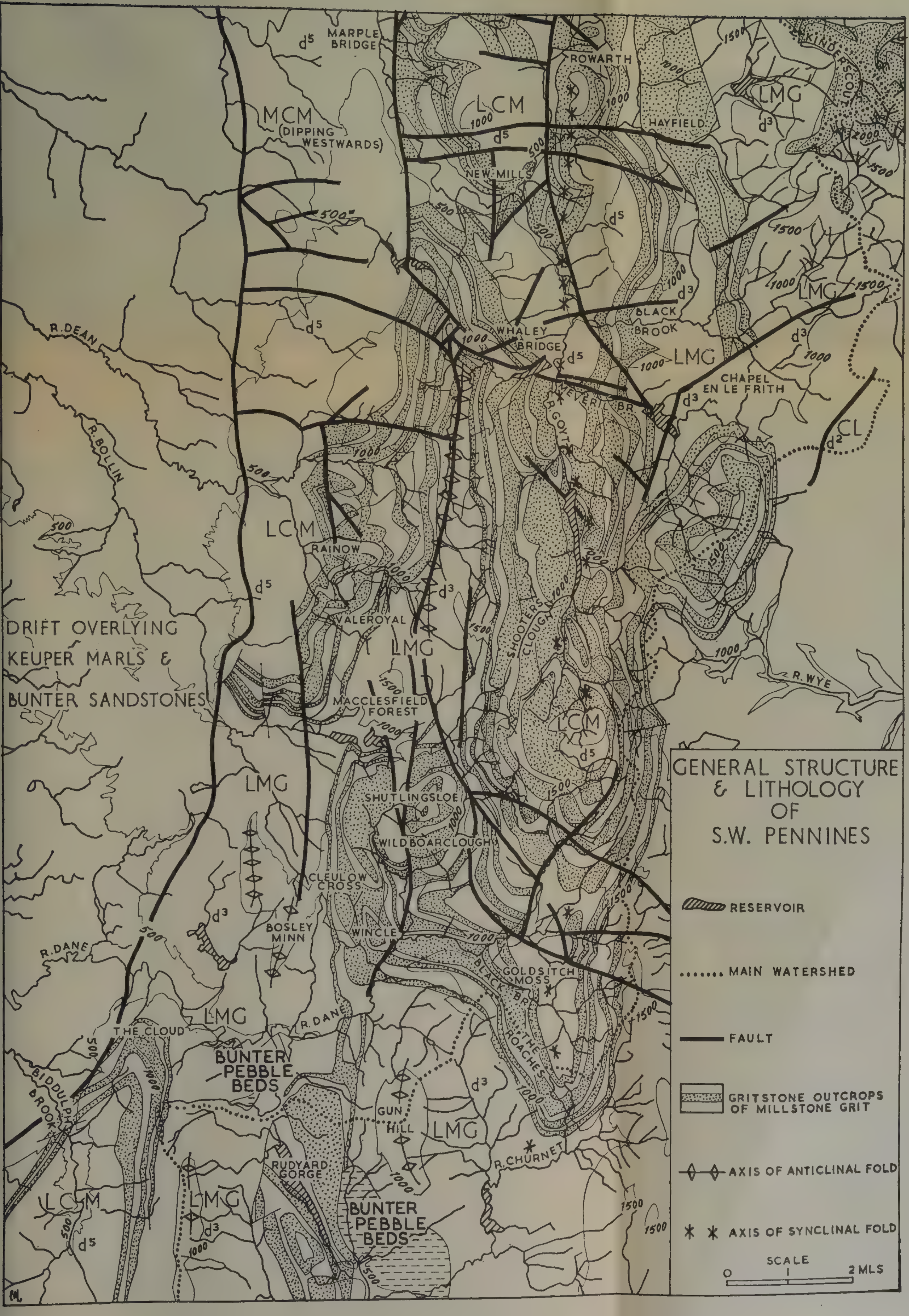

Fig. 2. General Structure and Lithology of SW. Pennines (based on Geological Survey maps)
d^3 L.M.G. Lower Millstone Grit Series (the other Millstone Grit series are indicated on the map)
d^5 L.C.M. Lower Coal Measures
M.C.M. Middle Coal Measures

[*To face p.* 24

follows the line of a syncline. The form of the valley is here much influenced by the dip of the major grit beds which, when stripped of the overlying shale, produce recognisable facets sloping steeply riverwards. Nevertheless, there are many convex breaks of slope not initiated by differential weathering but occurring at the intersections of older and newer valley forms. An attempt to show that these bear no simple relationship to the geological outcrops has been made in Fig. 4. In the Goyt system it is possible to trace the three series of benches upstream until, at heights of 1210, 1065 and *c.* 850 feet, they pass into short sections of flood-plain.

Moreover, corresponding benches can be seen in such tributary valleys as that of the Todd Brook, Shooters Clough and Wildmoorstone Clough. In contrast with the Goyt, the Dane flows through an area of great complexity. The distribution of valley benches is consequently very erratic; for in some localities the structure has favoured the preservation of benches, in others it has not. Moreover, in the Dane Valley below Bearda the solid rock is often so thickly covered with glacial drift that it is extremely difficult to determine the form of the sub-drift surface. The lowest series of benches in this area, therefore, is almost entirely obscured by drift. The Dean Basin also bears the clear imprint of the same sequence of events as that of the Goyt and Dane, but in the Bollin Valley only a few bevelled spurs bear testimony to the former river levels.

The similar physiographic histories of the Goyt, Dane, Bollin and Dean Valleys are consistent with an intermittently falling sea-level and yet there is no trace of ancient shorelines along the margin of the high ground. The probable explanation is that the shorelines lay farther west, and that the evidence for them has been destroyed by the relatively rapid erosion of the Triassic rocks of the Cheshire Plain. Today the series of benches which can be traced along the valleys in the Pennines end abruptly at the steep slope which marks the boundary with the Cheshire Plain. The fact that at the edge of the higher ground the corresponding benches are at different heights in different valleys supports the idea that the contemporaneous shorelines lay farther west. Thus, during the second phase of Table I the sea cannot have stood above 690 feet O.D. and the Dean, Bollin and Dane had at least seventy, sixty and sixty feet to fall after they left the upland.

3. THE RELATIONSHIP BETWEEN THE RIVERS AND THE GEOLOGICAL STRUCTURES OVER WHICH THEY FLOW

During the Hercynian orogeny the Millstone Grit and Coal Measures of the south-western Pennines were rucked into a series of complex folds along north–south axes (Fig. 2). The most important fold is the Goyt Trough which extends from near Hayfield in the north at least as far south as The Roaches. West of this major structure the folds are less continuous.

North of the Bollin Valley there is a faulted anticline of considerable amplitude breached by the Todd Brook. Between the Bollin Valley and the Rudyard Valley the strata have been folded into a series of domes elongated along north–south axes. Finally, south-west of the Rudyard Valley the northern end of the Potteries synclinorium is represented by the broad Biddulph syncline and the convergent Lask Edge and Astbury anticlines.

Both the Goyt and the Dane rise close to the axis of the Goyt Trough and so have to traverse a variety of structures before they reach the Permo–

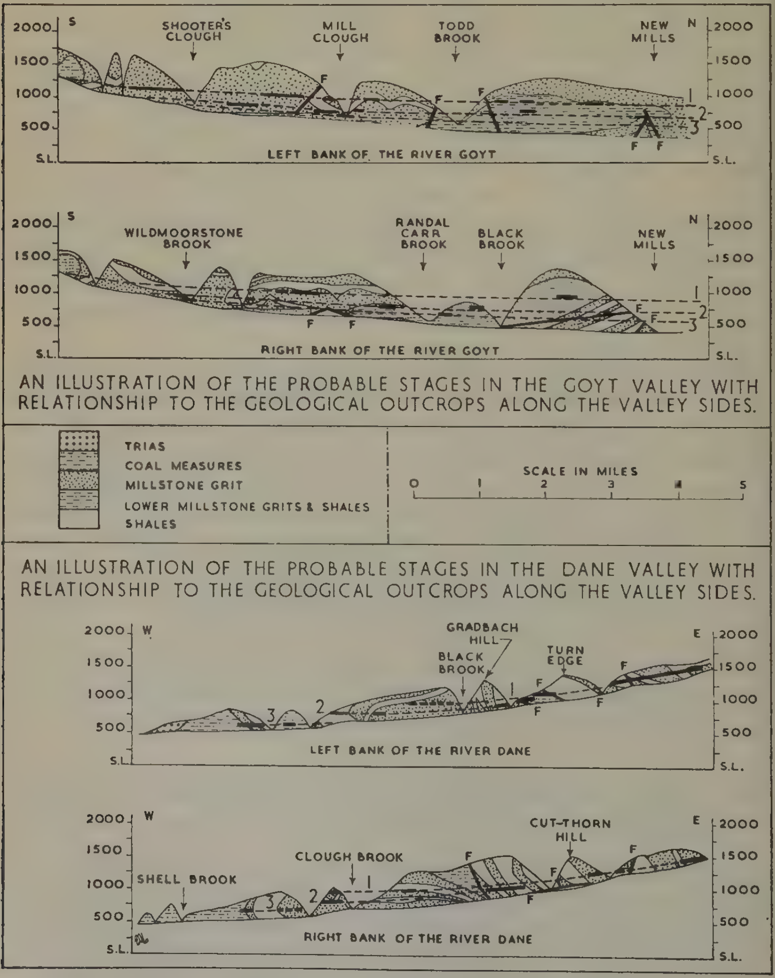
2000
1500
1000
500
S.L.
S
SHOOTER'S CLOUGH
MILL CLOUGH
TODD BROOK
NEW MILLS
N
F
1
2
3
LEFT BANK OF THE RIVER GOYT
WILDMOORSTONE BROOK
RANDAL CARR BROOK
BLACK BROOK
RIGHT BANK OF THE RIVER GOYT
AN ILLUSTRATION OF THE PROBABLE STAGES IN THE GOYT VALLEY WITH RELATIONSHIP TO THE GEOLOGICAL OUTCROPS ALONG THE VALLEY SIDES.
TRIAS
COAL MEASURES
MILLSTONE GRIT
LOWER MILLSTONE GRITS & SHALES
SHALES
SCALE IN MILES
0
1
2
3
4
5
AN ILLUSTRATION OF THE PROBABLE STAGES IN THE DANE VALLEY WITH RELATIONSHIP TO THE GEOLOGICAL OUTCROPS ALONG THE VALLEY SIDES.
W
E
GRADBACH HILL
BLACK BROOK
TURN EDGE
LEFT BANK OF THE RIVER DANE
SHELL BROOK
CLOUGH BROOK
CUT-THORN HILL
RIGHT BANK OF THE RIVER DANE

Fig. 4. An illustration of the Probable Stages in the Goyt Valley and Dane Valley

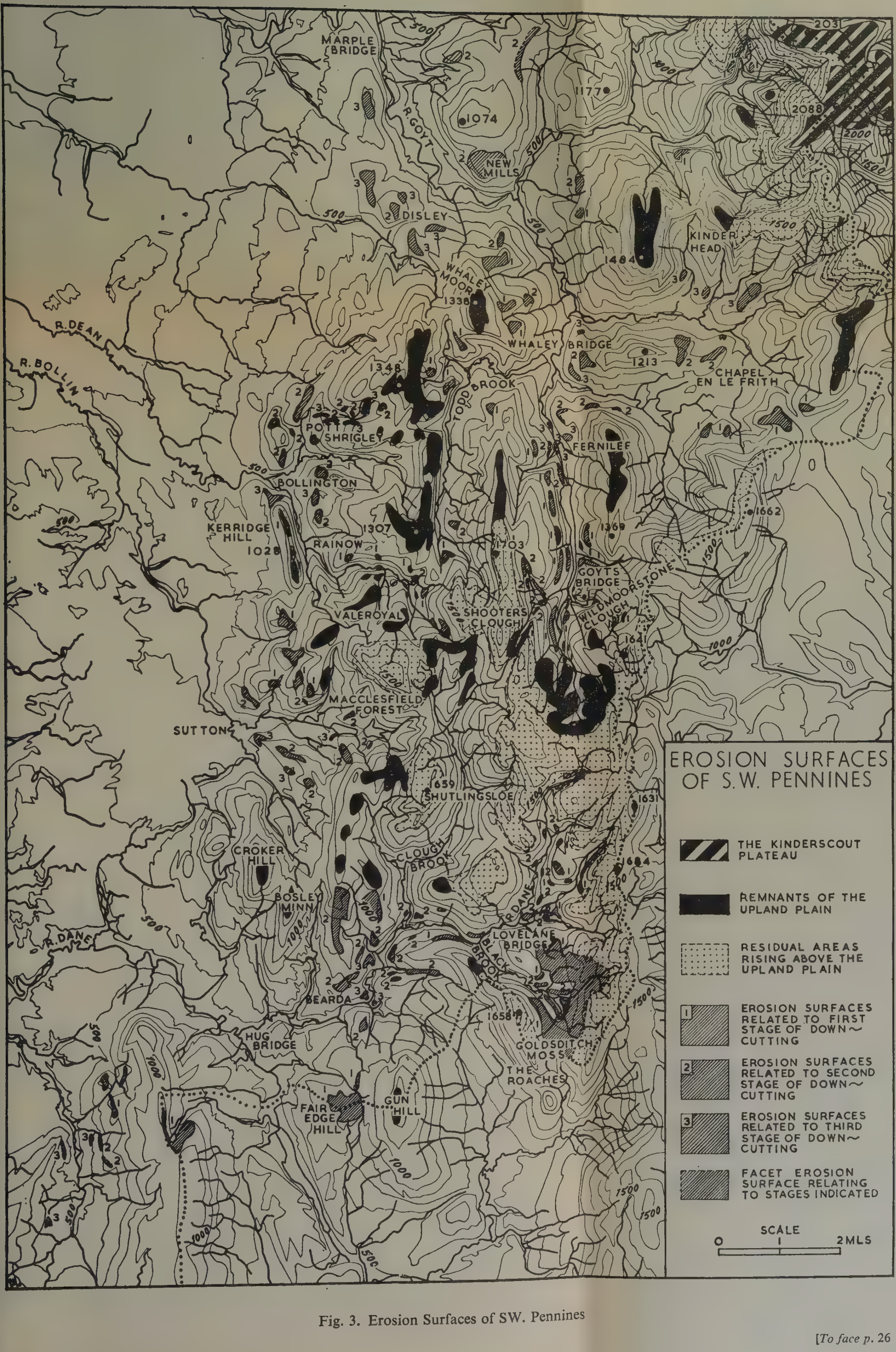

Fig. 3. Erosion Surfaces of SW. Pennines

[*To face p.* 26

Triassic rocks of the Cheshire Plain. In considering the relationship between the river courses and the geological structures two major problems are encountered. Firstly, as tectonic crumpling and dislocation occurs not along lines but along zones, exact coincidence of a subsequent river course and a fault or fold axis as depicted on a map is not to be expected. Yet the latitude to be allowed a stream which may still be regarded as subsequent poses a nearly intractable problem which requires for its solution a minute knowledge of both geological structure and stream history. In its lower course the Black Brook which joins the Goyt near Whaley Bridge flows through two prominent grit ridges. The stream is here paralleled by a fault which outcrops about quarter of a mile from the river and causes a striking discontinuity in the run of the grit ribs. In such circumstances a stream is probably best regarded as subsequent.

The second and obviously related problem concerns the lack of adjustment sometimes shown by streams which in an earlier phase were structurally guided. From the form of its valley, it is clear that Shooter's Brook, a left bank tributary of the Goyt, has shifted uniclinally down-dip along a shale-grit junction. Today, however, the stream has succeeded at a number of points in incising itself to a depth of five feet into the joint system of the underlying grit. Short of a cataclysmal flood and the cutting of a completely new channel further down-dip, the stream will in future be anchored in the grit outcrop. This process of 'disadjustment' seems to be responsible for a number of instances in which rivers flow along the outcrop of thick grit beds. The most important involves the Dane itself, which, for one and a quarter mile, flows along the strike of the massive Roaches Grit. Since a corresponding course is followed by the Black Brook immediately above its confluence with the Dane, the Roaches Grit outcrop is marked for two miles not by an escarpment but by a valley. Moreover, the valley follows the curvature of the strike so exactly that a fortuitous correspondence seems highly unlikely. It appears that the Dane and Black Brook were at one time guided by the shale overlying the Roaches Grit, but that they have since incised themselves into the more resistant bed.

When, with the above considerations in mind, the courses of the Goyt, Dean, Bollin and Dane are examined, the noteworthy fact emerges that there are only very few miles of river which cannot, with reasonable confidence, be referred to structural guidance. The Goyt from its source to Whaley Bridge follows approximately the axis of the Goyt Trough. Downstream it swings across the western limb of the fold along a NNW.–SSE. fault zone, and from New Mills to Marple, despite masking of the tectonic control by glacial diversion (Rice, 1957), follows the faulted crest of the Todd Brook anticline. Finally, below Marple Bridge it crosses the lower ground of the Coal Measures, still without flowing transverse to any major structure. The tributaries of the Goyt appear equally related to geological structure; the smaller have generally etched valleys along shale outcrops, the larger along faults.

Although the Dane headstream escapes from the Goyt Trough along a short section unrelated to any known structure, the river soon begins to follow the line of a fault trending NNE.–SSW. Where it leaves this fault it follows the outcrop of the Roaches Grit and provides the example of disadjustment described above. At Wincle the river again turns along a major fault, before reverting to its westerly direction and traversing the zone of periclines. The structures in this latter zone are extremely complex with rapid change of dip, and although the Dane may cross a number of fold axes it also tends to follow the tectonic cols between the major domes. Similarly, the courses of almost all the Dane tributaries are clearly guided by geological structure.

For just over one mile in its upper reaches the Dean cuts across a number of grit ribs without known structural guidance. But at Rainow it turns northwards along a shale vale and finally leaves the Pennines where the Kerridge Hill scarp is broken by a fault. The site of the Bollin Valley breach in the Pennine front may also be related to local complex tectonics (Pocock *et al.*, 1906, 56).

In summary it may be said that in a distance of twenty miles along the south-western flank of the Pennines there is not one incontrovertible case of a stream transecting a major fold axis without structural guidance. The remarkable contrast with the eastern flank of the Pennines is immediately apparent from an inspection of the maps prepared by Sissons (1954).

4. A DENUDATION CHRONOLOGY FOR THE SOUTH-WESTERN PENNINES

The study of the drainage pattern tends to confirm the view expressed earlier that the Upland Plain is not a marine platform. It seems most unlikely that a series of superimposed consequents initiated at the level of the Upland Plain could have been so disrupted as to be almost completely obliterated. Moreover, there is very little evidence of river capture below the level of the Upland Plain. Only two possible instances were noted. The Valeroyal tributary of the Dean (Fig. 3) probably continued westwards on to the Cheshire Plain until, at a height of about 900 feet, it was captured by a feeder of the middle Dean. More recently some Meveril Brook headstreams near Chapel-en-le-Frith may have been diverted by the Black Brook, but this is competition between subsequents, itself a testimony to the advanced adaptation to structure. In 1954 (Sissons, 322–3) listed seven features as characteristic of drainage initiated on marine platforms; not one of the seven is characteristic of the drainage in the south-western Pennines.

It is concluded, therefore, that the present drainage pattern is considerably older than the Upland Plain; but since it is composed almost entirely of subsequent streams, it is difficult (if not impossible) to reconstruct any earlier pattern. It is none the less instructive to consider possible reasons why the original pattern has been so completely replaced. One obvious factor is the contrasting resistance of the shale and grit beds which will tend to encourage the growth of subsequents. Yet an equal contrast is found on the eastern side of the Pennines, and it therefore seems reasonable to ascribe the difference between the eastern and western sides of the Pennines to a difference in geomorphological history. It may be that the consequents of the south-western Pennines were superimposed on the Carboniferous rocks at an earlier stage than those of the eastern Pennines, thus allowing more time for their disruption. An alternative view is suggested by Linton's thesis (1951) that, although over most of Britain the Chalk cover dipped eastwards, west of the Pennines, it was warped down in a huge monoclinal fold or subjected to immense faulting. If this were so, two sets of consequents would originate, one flowing to the east with gentle gradients, the other to the west with steep gradients. Rapid divide migration would ensue, with the west-flowing streams constantly enlarging their catchment area. Once the Chalk cover was breached, the headward extension of the west-flowing streams would be very closely guided by structure, and this might in turn involve the almost complete replacement of superimposed consequents by subsequents along the south-western flank of the Pennines (cf. Holtedahl, 1952).

Although the foregoing paragraph is highly speculative, it indicates how a study of the western edge of the Pennines may assist in elucidating the geomorphological evolution of the whole upland. Whilst the present paper is concerned with too small an area to permit the formulation of a complete denudation chronology, there is much evidence to support the following conclusions.

The drainage pattern of the south-western Pennines had already evolved to near its present state when there was a prolonged stable period of subaerial erosion during which the landscape advanced to a stage of late maturity. A widespread but undulating surface was eroded across grit and shales alike, and is today found at a height of 1200–1400 feet. After this stable period there was intermittent uplift with three short stillstands represented in the valleys of the south-western Pennines by three series of benches. All this time the coastline lay west of the Pennines and so there is no trace of marine platforms.

Finally, the relationship of the above surfaces to those which have been mapped elsewhere in the southern Pennines may be briefly considered. In the valleys of those streams which drain eastwards from the major water parting to the Derwent, benches can be found which pass downstream into the (Peak District) Upland Surface (Linton, 1956; Johnson, 1957). This has a lower limit of *c.* 900 feet and marks the end of the first phase of incision below the Pennine summits. It thus occupies a comparable position to the Upland Plain of the south-western Pennines which lies well below the level of such summits as Kinderscout and Axe Edge. Despite the discrepancy in the height of their lower limits (which may be quite simply explained by the absence of rocks of sufficient resistance in the Cheshire Plain to preserve such a high-level erosion surface) the (Peak District) Upland Surface and the Upland Plain have much in common, their general similarity of form and especially their widespread extent being strongly suggestive of contemporaneity in development and later dissection. It is also noteworthy that they both mark the last occasion on which the relief of their respective areas was reduced to that of late maturity.

ACKNOWLEDGMENTS

The writers wish to acknowledge the help given them by Professor P. R. Crowe, Professor D. L. Linton, Dr. I. M. Simpson and Mr. B. J. Taylor, in the preparation of this paper, and to thank Miss E. A. Lowcock for her careful work in redrawing the illustrations. They are also grateful for a grant towards field expenses from Manchester University.

APPENDIX

A note on the construction of Figure 4

In these diagrams drawn for the Upper Goyt and Dane Valleys an attempt has been made to demonstrate the relationship between the valley benches and the geology. The diagrams were constructed by imagining a plane running along the centre of the valley on to which the geological outcrops and the valley benches were projected. The instances in which this method can be successfully used tend to be restricted by two sources of distortion: (a) unless the valley is straight and steep sided considerable distortion will occur at bends; (b) the need for vertical exaggeration will so increase apparent dip of the rocks as to destroy the pictorial illusion. In the case of the Goyt Valley neither of these difficulties is encountered since the valley is relatively straight and the dip of the rocks at right angles to the river direction. The Dane reconstruction is less successful but the relationship of benches and geology appears to be capable of satisfactory demonstration by this method.

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Longitudinal Ripples in the Upper Tunbridge Wells Delta, Kent, and their Probable Mode of Origin

by F. G. BERRY

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BSTRACT: Structures interpreted as longitudinal ripples similar to those found in arine sediments are described for the first time from non-marine sediments and a ypothesis explaining their formation is developed. Ripples of this type are likely to rm during the transport of small quantities of sand by relatively fast currents across ud-flats.

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1. INTRODUCTION

ONGITUDINAL RIPPLES cut in fine-grained sediments (van Straaten, 1951) re distinctly rare in non-marine environments. In fact, apart from a brief eference to giant forms in rivers (Chenoweth, 1955) there appear to be no revious reports or descriptions in the literature. The occurrence here escribed yields details of their structure which endorse the observations of an Straaten as to their erosional origin—although it is recognised that imilar forms may arise by accumulation (van Straaten, 1953). Either type ay be subsequently propagated by conformable sedimentation (Kelling, 958).

2. LOCATION AND GENERAL STRATIGRAPHY

A degraded section at Sandhole Farm, five miles west of Tonbridge, ent (509549),[1] exposes the junction between a fine-grained laminated andstone and gray silty clays above. These occur well up in the third Vealden megacyclothem (Upper Tunbridge Wells–Lower Weald Clay) ostulated by Allen (1959) and confirmed locally by the writer's detailed mapping. They are some distance below the abundant red-mottled silty clays which are customarily taken as the base of the Weald Clay. To the west of Sandhole the sandstone passes beneath the Eden alluvium and is not known to reappear. To the east it is breached by a small tributary of the Eden, but reappears in the vicinity of Moorden Farm (523459). Between this point and Redleaf Wood it begins to thicken and form natural bluffs. It disappears from outcrop within the next hundred yards. The available field evidence strongly suggests that the sandstone is a linear deposit of 'lenticular' cross-section, following the local strike trend where exposed and truncated at Redleaf Wood by a change in topography.

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[1] All Grid References lie within the 100 km. square 51 (TQ) 54.

The silty clays above—about thirty feet thick at Sandhole—pass vertically and laterally into flaggy siltstones interbedded with unlithified silt, silt–clay grades and waxy clays, usually with, *inter alia*, some fine sands. Much of this type of sediment is currently regarded as levee alluvium.

3. DESCRIPTION OF THE SEQUENCE

The passage from the sandstone to the silty clay above is associated locally with retreat structures similar to those described by Allen (1949, 1959) from the Ashdown–Wadhurst and Lower Tunbridge Wells–Grinstead junctions (Fig. 1). The basal sands (a) are followed by a series of oscillation ripple systems (b) derived from current-formed precursors, built,

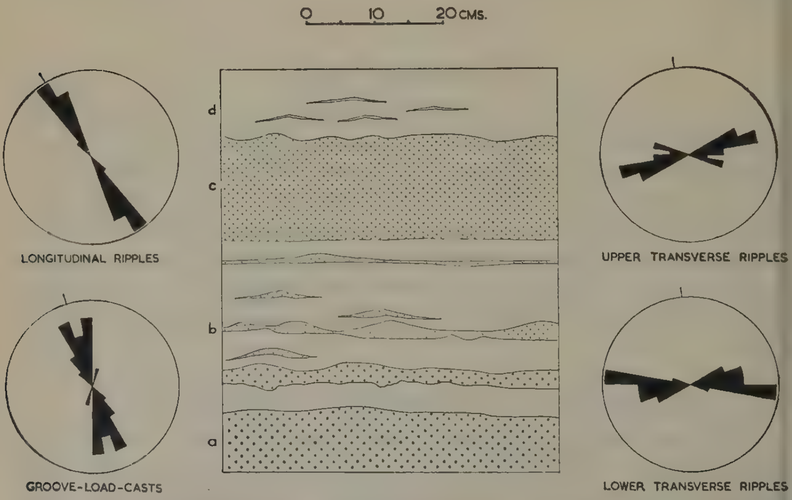

Fig. 1. The 'retreat' sequence at Sandhole Farm, Kent, with orientation histograms of the lower transverse ripples from 'b', groove-load-casts from the base of 'c', longitudinal ripples from the upper surface of 'c' and upper transverse ripples from 'd'. Coarse stippling = very fine sands, fine stippling = silts, white = clay-silt grades

near the base, of very fine sand and, higher up, of coarse and fine silts. They are separated by clayey silts. This ripple sequence is succeeded by a relatively thick, very fine sand–siltstone (c) the base of which bears a variety of groove-load-casts. At one point a diminutive piercement structure formed by the injection of the underlying clayey silts has bent these up into a near vertical position.

The upper surface of this unit is furrowed by the system of longitudinal ripples.

The siltstone is followed abruptly by more clayey silts (d) containing only the faintest traces of silt-crowned transverse ripples, some of which occur only two or three centimetres above the longitudinal ripples.

No pebble or rootlet beds have so far been observed and it is doubtful whether these retreat structures, in the absence of thick clays of Wadhurst or Grinstead type, signal the onset of a widespread lake transgression.

The orientations of oscillation ripples above and below the siltstone (c) and the top- and bottom-surface structures are shown in circular histograms in Fig. 1. It would appear from their arrangement that the wave oscillations were normal to the current trends that produced the longitudinal structures and that the strikes of the current-formed precursors have been preserved. The arithmetic means of these trends are 355, 343, 328 and 351 degrees, in ascending order. Whether currents moved predominantly to the north or south is not clear, for ripples propagate in opposed directions and groove-load-cast bifurcation is also inconsistent.

4. DESCRIPTION OF THE LONGITUDINAL RIPPLES

In plan the ripples show an essentially regular and parallel arrangement, with an occasional bifurcation. Fig. 2 shows a series of transverse sections through one of these bifurcations. The surface profile is smoothly rounded but less regular than that described by van Straaten (1951). The wavelengths of this surface vary between 1.5 and 7.0 centimetres (average 4.2 centimetres) with amplitudes ranging from 2.0 to 0.3 centimetres (average 0.7 centimetres). These yield ripple indices of from 18.0 to 3.5.

The smooth profile is due to a coating of fine silts masking an underlying and more irregular surface (indicated in the figure by heavy lines). In all cases where the original, horizontal laminations in the siltstone can be observed it is cut by this surface, which is clearly an eroded one.

The silt covering conceals a variety of slots and shallow, round or flat-bottomed grooves in the troughs. The base of the infilling then usually consists of very fine, sharp sands along some part of the grooves. The fine silts above the sands generally show discordant bedding, but the laminae ometimes curve upwards at steep angles near the sides (Fig. 2c). This may be due, in part, to compaction, though some sections (a) indicate that it is not. Two distinct layers of sand are not infrequently present, separated by fine silts. In longitudinal section the infilling appears to be level bedded. The fact that some of the grooves have vertical walls and are occasionally undercut (a) suggests that the muds were partially consolidated before the grooves were cut. One small slump has been observed in the fine silt covering.

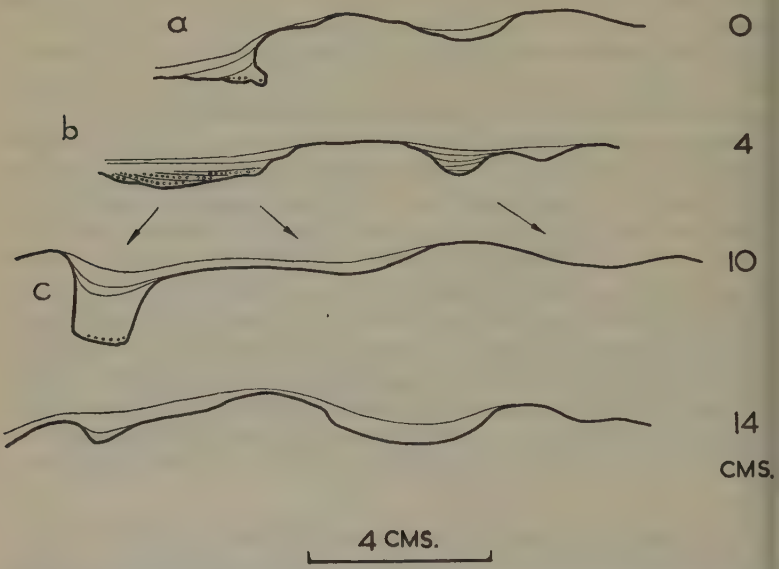

Fig. 2. Transverse serial sections through the bifurcation of a longitudinal ripple. (Distance between the sections is indicated at the side)

5. SUGGESTED MODE OF ORIGIN

By analogy with aeolian conditions (Bagnold, 1942, 177) it seems that strong currents may favour transverse instability. Casey (*in* Bagnold) found that longitudinal strips of material were deposited under water only when the velocity of flow was raised *above* a certain level and was associated with regular transverse elements These facts apply primarily to accumulative processes and will no doubt lead to an explanation of the way in which certain types of longitudinal ripples are formed; but linked with the foregoing description an hypothesis concerning the origin of indubitable erosion-formed longitudinal ripples may be developed

Van Straaten (1951) recorded the occurrence of transverse current ripples in sand adjacent to longitudinal ripples in muds Both were clearly stable under conditions of similar current strength—implying a considerable difference in the nature of the bottom flow over the two surfaces. This is evidently related to bottom roughness and the power of the individual grains to move and interact with the boundary layers (Inman, 1949). Over sands current flow is retarded (whether ripples are present or not) while over adjacent muds, current flow is, relatively speaking, accelerated.

However, if small quantities of sand become washed over the muds from adjacent areas (or, alternatively, are winnowed from them) bottom flow over the muds will tend to move in preferred paths (since flow over the portions covered by the sands is retarded) so drawing streamers of sand along in the direction of the current. These will remove particles beneath them into suspension by traction and impact, assisted by greater bottom roughness, and thus promote the formation of grooves. Current speed must eventually diminish and the sand grains will accumulate then in the bottoms of the grooves—generally where they are widest (Fig. 2b). At this point the oscillatory movements due to waves become effective and impart slight to-and-fro movements to the sand grains which locally deepen the grooves. Ultimately, the finer material begins to settle from suspension to complete the filling. At this stage a local resurgence of currents or wave oscillations may form double layers of sand and cut into the groove filling to form the curved laminae (Fig. 2a).

A more general form of scour will continue so long as material remains in suspension—gently rounding the profiles of the grooves.

It is clear that longitudinal ripples can form only when the supply of sand being carried over the muds remains below some (unknown) proportion, appropriate to current strength; for if it should exceed this amount, bottom flow will be retarded and transverse ripples form over an unbroken blanket of sand in which preferred paths cannot be established (this situation should not be confused with the formation of certain groove-casts and similar structures found on the base of thick-graded beds and possibly due to a similar mechanism). The supply of sand will also influence the wave-length (mean spacing)—again in relation to current strength, since it controls the width and thickness of the slow streams and hence their ability to resist longitudinal dissection by faster flow.

6. DISCUSSION

Contrary to Kelling (1958, 129) there seems to be nothing in the presence of flat-bottomed grooves and uniform parallel ridges to invalidate a scour hypothesis operating in the manner and conditions outlined above. By analogy with aeolian conditions again (Bagnold, 1942, 232) a uniform and parallel arrangement is to be expected in generally uniform conditions of flow over level mud-flats; the uniformity is related to secondary transverse circulation or the probability that eddy zones between fast and slow streams are of similar competence, while their parallel arrangement is due to the absence of small bumps or other obstructions that could divert the streams.

The part played by wave-height is uncertain. Van Straaten (1951) found a correlation between (a) wave-height (b) wave-length of the longitudinal ripples and (c) depth of water. That depth of water should be related to wave-length of the longitudinal ripples seems very probable since depth of water is also related by van Straaten to current velocity—an integral part of the process of erosion. It seems significant that the smaller wave-lengths he reported occurred in the region of fastest currents, where the sand streams might well be attenuated and hence split into thinner and more frequent streams, so producing longitudinal ripples closer together.

Regarding the source of the currents which formed the longitudinal ripples, much depends on the correct interpretation of the adjacent sediments. The basal sands may represent, as their linearity suggests, the bar underlying a minor distributary pass and the local grey clays, accumulations within an abandoned channel receiving intermittent flood waters. If the currents are attributed to crevassing the infrequence of this type of longitudinal ripples is surprising, since the sediments requisite in their formation constitute the bulk of the Hastings Beds. It seems more probable that they are passed over as ripples of normal type.

7. CONCLUSIONS

(a) Longitudinal ripples superficially similar to those of van Straaten (1951) are described from a non-marine and presumably tide-less environment.

(b) The ripples were formed by erosion.

(c) Evidence is reviewed which supports the hypothesis that ripples of this type are most likely to form on level mud-flats when a small quantity of sand is being transported across them by fast currents.

(d) The parallel and generally uniform appearance of the ripples is no more than would be expected in the conditions of uniform flow that may occur over mud-flats.

(e) Wave-length may be related to strength of current flow and quantity of sand available.

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An Examination of Otmoor and the Adjacent Areas of the Oxford Clay Lowland

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ABSTRACT: The Oxford Clay lowland, east of Oxford, is drained by the River Ray, a tributary of the Cherwell. The river flows across two distinct basins, an upper basin near Marsh Gibbon and a lower basin known as Otmoor. The paper describes the results of some examinations of the basins, conducted in an attempt to find and map the alluvium which is generally thought to exist on Otmoor.

Various methods were used, historical, topographical and geological, in the study. The principal conclusion reached is that Otmoor is not an alluvial basin but a normal clay lowland that suffers from poor drainage. The Ray has not always flowed through Otmoor and its development has probably gone on independently of events affecting the river.

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1. INTRODUCTION

NORTH-EAST of Oxford, the Oxford Clay lowland extends for some twelve miles before reaching the edge of the glacial drift deposits, which forms a marked boundary. It is drained by the River Ray, a Cherwell tributary. The Ray flows across two distinct basins, separated by a constriction caused by the northward outlier of the Corallian; Arncott Hill, and the small ridge at Ambrosden. The lower basin, Otmoor, is two miles above the Ray–Cherwell confluence at Islip; the upper basin is around Marsh Gibbon. The research behind this paper has been carried out principally on Otmoor but reference has also been made to the upper basin.

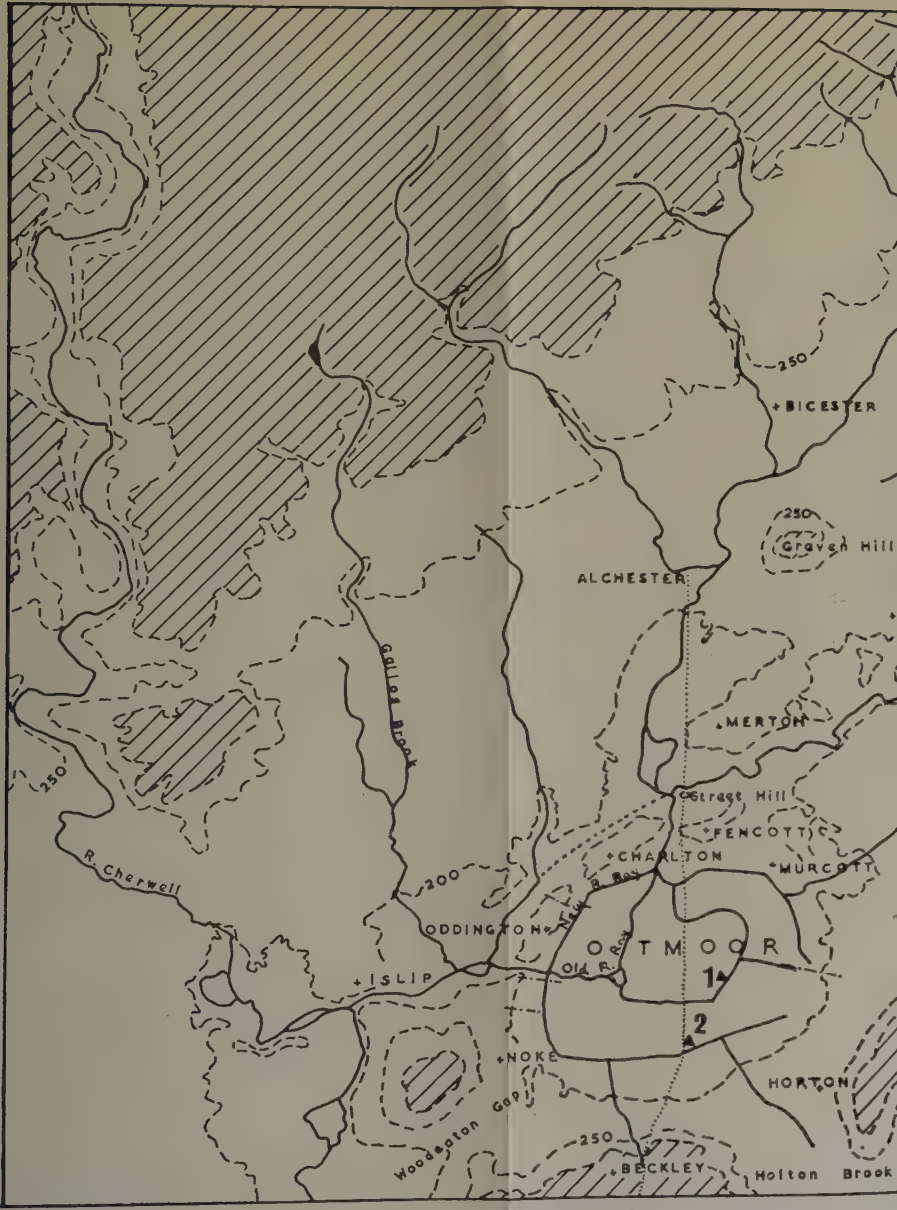
Otmoor lies some seven and a half miles from Oxford but, despite its proximity to the city, it has been a blank in the otherwise fairly well-documented geology and geography of the Oxford region. This may be attributed to the wet conditions which persist there through much of the year, and to its use from 1933 to 1958 as a bombing range by the Royal Air Force. Otmoor is regarded locally as being rather remote, primitive and mysterious, contrasting with the life of its surrounding area. The late Dr. W. J. Arkell (1947, 247) described it as an outstanding occurrence of old alluvial deposits, peats and humus which offered a rich field for future research.

Otmoor has an area of some four square miles and lies across the lower course of the Ray about a mile upstream from the gorge at Islip, where the stream cuts through the Cornbrash ridge to join the Cherwell. The moor is developed on Oxford Clay and lies between 200 and 193 feet O.D. It is surrounded on three sides by higher ground but is open on the east and north-east, where it is entered by the River Ray. The hills to the south and west are composed of Calcareous Grit, Coral Rag and Arngrove Stone, with patches of Kimmeridge Clay. These western hills belong to the Wheatley fault-zone and their structure has been described by Arkell (1942, 199–203). The northern boundary is formed by low periclines of Cornbrash and Kellaways beds. These periclines rise to only slightly over 210 feet O.D. but form a distinct boundary to the moor. They are continued in the upper Ray basin but do not form the limit of the area. The upper Ray basin is bounded by the Oolitic rocks to the north and upper Jurassic beds to the south. The eastern boundary is the Boulder Clay.

The earlier Geological Survey maps showed Otmoor as an area of Oxford Clay, but the later maps, which include downwash and local material as 'alluvium', show it as a considerable expanse of this material. Arkell accepted this but modified the Survey's boundaries (Arkell, 1944, map). The principal object of our work has been to find this alluvium, its sources, and to try to map its extent; to examine the adjacent parts of the Oxford Clay vale and the drainage system in general. By so doing we hoped to be able to link our results with those of workers in the upper Thames area. This paper is not the summary of a completed task so much as a progress report.

The River Ray has a total length of some fourteen miles. The overall gradient of the stream varies with its tributaries. The main stream from Grendon Underwood has a fall of some sixty feet. The tributary from Stratton Audley, which joins the river at Fencott, falls eighty feet, but those coming from the edge of the boulder clay fall only some forty feet from their source to the junction of the Ray with the Cherwell. In all cases the major part of the fall occurs in the upper part of the course.

The majority of the Ray tributaries converge in the Marsh Gibbon basin. This basin is similar to Otmoor, but has more pronounced relief. The Ray leaves its upper basin at 203 feet O.D. and enters Otmoor, six miles downstream, at 198 feet O.D. Erosion is concentrated in the upper part of the system. Below Three Points most of the stream's power is absorbed in the transportation of material and the mechanical factors of flow. Any alluvial material encountered in the basin, except for local washdown, may be



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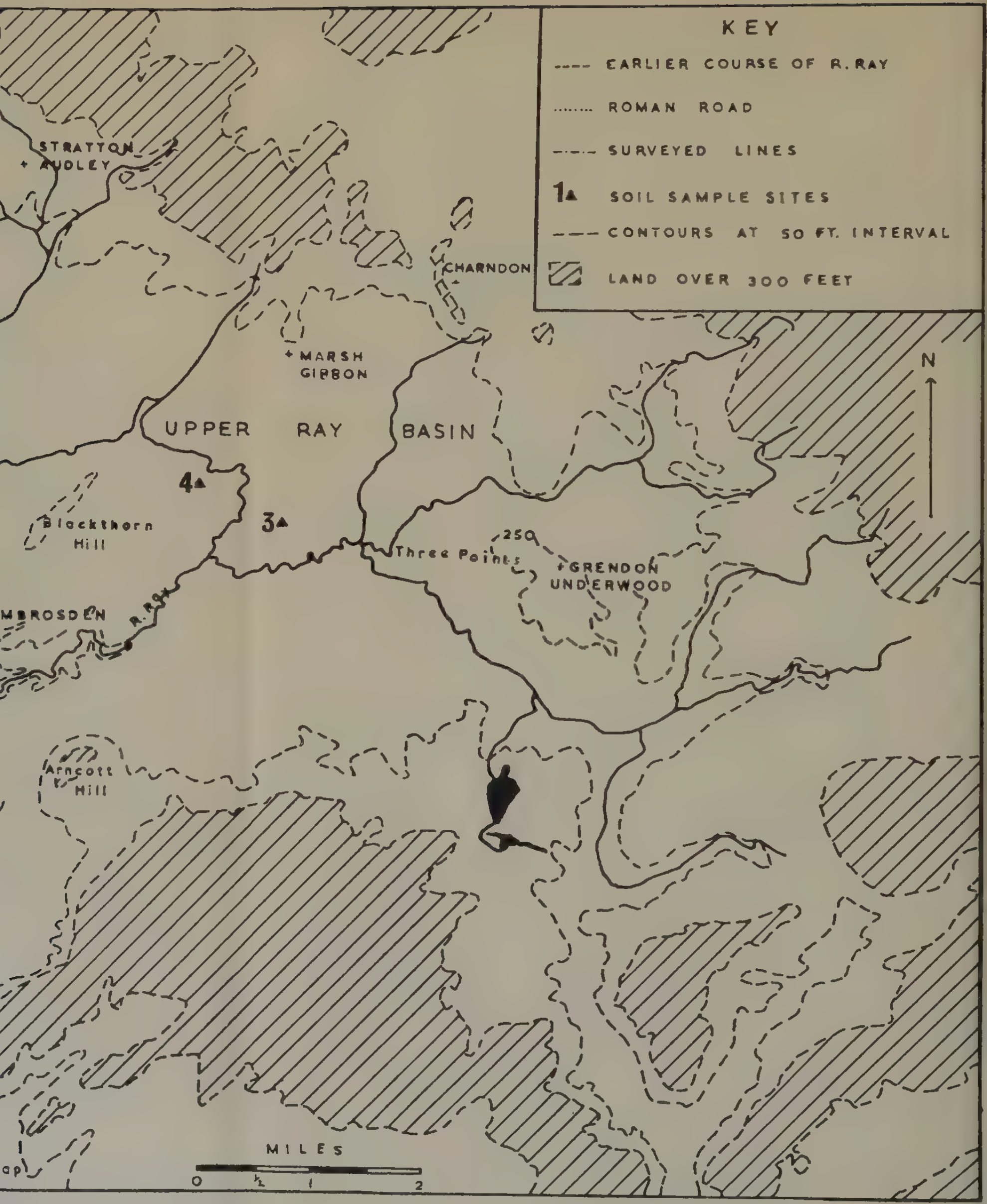

oor and the upper Ray basin

expected to be derived from the Oolites, the Corallian, the Oxford Clay, or from drift-derived flints, quartzites and other erratics. The passage of melt-water through the Ray system has not yet been proved (Lampugh, 1920, lxxiv).

Various theories have been advanced as to the origin of Otmoor, some more plausible than others. One heard, but not read, was that it was a meteorite crater! The general theory suggests that it has developed by the ponding back of the Ray, possibly as a lake, above its little gorge at Islip, and the gap at Oddington (Arkell, 1947, 225). This ponding back has been attributed to Cherwell flood-waters and their deposits (Beckinsale, 1954, 31).

We have approached the problem in four ways. Firstly, historical evidence and aerial photographs were studied. Secondly, some levelling was carried out in conjunction with a study of bench-marks, both on and around Otmoor. Thirdly, soil augers were used on the moor and in other parts of the basin to determine the nature of the soil and subsoil. Fourthly, mechanical analysis of selected soil samples was made.

2. HISTORICAL EVIDENCE

The Roman Road from Alchester to Dorchester (Oxon) crosses Otmoor between Beckley and Fencott (Fig. 1). It is built up on a causeway of Corallian rubble with a surface dressing of finer material. It is in a fair state of preservation and has not been extensively rebuilt. That it is still clearly defined contrasts strongly with the Saxon causeway built at Binsey on Thames Alluvium which now lies six feet below the general level of the land (Hussey, 1841). North of Fencott the line of the road is less distinct, particularly where it crosses the area of alluvium derived from the Oolites, near Merton.

The drainage of Otmoor was much modified in the 1820s by attempts to improve the moor (Marshall, 1944, 215). The work was carried out following the reports by J. Davis (1795) and Arthur Young (1809) on the state of agriculture in the county. Both drew attention to the need to drain Otmoor, which, they thought, would provide good pasture land, if it were properly enclosed. Young said: 'I searched and inquired for bog and peat, but have reason to believe that there is no such thing upon it [Otmoor], or if there be, in very small proportion. To the eye it is level, but there is inequality enough to make waters draw down to one part more than the rest, so that it is very wet in October' (Young, 1805, 227).

Despite the drainage works which give the moor its striking appearance on the topographical map, the wet conditions persist today, probably because the drainage plan was never completed. The New Ray, which was dug along the north-west edge of the moor, carries water collected by the upper stream. The moor itself is drained by the remnants of the Old Ray and a system of artificial ditches. Aerial photographs suggest that Otmoor was drained by a fan-like system converging on the gap in the Cornbrash at Oddington and linking with the Old Ray.

3. LEVELLING

Representative sections of the slopes bounding the moor were levelled and profiles drawn. They showed no traces of a bench or series of benches which might have been associated with a lake. The composite section is shown in Fig. 2.

The study of the distribution of Ordnance Survey bench-marks gives the impression that Otmoor may be lowest in the south-west. The majority of these bench-marks are on gate-posts leading off roads which surround the moor and which are embanked above the general level. The normal drainage slope is towards the line of the Old Ray.

4. AUGERING

Systematic examination of the moor surface and other parts of the clay basin was carried out by augering. Three-foot and five-foot corkscrew-type augers were used; these were adequate to penetrate the gley horizon and to determine if alluvium were present.

The majority of auger profiles on Otmoor have been very similar. The top four to nine inches, below a humic surface horizon, is a brown loam; this is underlain by brown or yellow mottled clay which passes down into grey clay. In some cases the clay has a platy structure and there are occasional bands of calcium carbonate.

TABLE I. **A Typical Otmoor Soil Profile**

	HORTON (42/579142)
Depth in inches	Description
0– 6	Dark brown peaty loam
6–15	Yellow brown clay
15–24	Yellow–grey mottled clay
24–30+	Dark grey clay.

Generally, soil colour changes occur within a few inches and in some cases there is distinct banding. Horizon differentiation was not possible owing to the compaction which occurs with a corkscrew auger.

A certain amount of downwashed material from the Calcareous Grit was encountered near Noke. Some sand and grit was found in an old stream channel but another similar channel was choked by a clay with a high vegetable matter content. A sample of this was ignited and the material lost represented 40% of the sample dry weight. Otmoor is almost entirely covered by clay-derived soils. The high vegetable matter content may be attributed to the damp conditions which stimulate plant growth and retard its decomposition. We have found no peat as such, but we have seen substances which, if dried, might provide fuel.

The Woodeaton and Holton Brook gaps leading out of Otmoor have been examined. They have been regarded as the result of spring sapping (Beckinsale, 1954, 32), and theories that they were overflow channels discounted (Sandford, 1926, 144–5). This method of formation is acceptable for the Holton Brook gap, but auger holes and a trench in the Woodeaton gap revealed water-worn gravels containing flints and quartzite pebbles, with some material apparently slumped from the valley sides. This would suggest that the gap was used by a stream heading in the east of the clay vale.

Between Fencott and Merton there are extensive areas of more sandy material which is probably Ray alluvium. The hamlets of Fencott and Murcott stand on this better drained material overlying the Oxford Clay. Alluvium is also encountered in the broad valley to the north of Charlton and Oddington, between the periclines and the main Cornbrash outcrop, and links with the Gallows Brook gravels near Islip (Sandford, 1927, 142). Associated with this alluvium is a discontinuous gravel terrace at about 205 feet which runs back from Oddington to Merton.

The possibility of the diversion of the Ray near Merton, by the more strongly flowing Stratton Audley tributary stream, southwards between the Charlton and Street Hill periclines along the line of a transverse fault or other weakness, has been considered. The small terrace is only slightly above flood-plain level, but diversion appears to have occurred subsequently to its formation. Thus it is probable that the diversion of the main Ray drainage through Otmoor is a comparatively recent feature.

Some parts of the upper Ray basin have been examined. The auger holes show that there is a considerable amount of silt and sand intermixed with yellow clay. The soil profiles differ noticeably from those of Otmoor in colouring, texture and vegetable matter content. The upper and lower Ray basins appear to differ importantly.

5. MECHANICAL ANALYSIS OF SOIL SAMPLES

The mechanical analyses of soil samples from Otmoor and Marsh Gibbon were done by the Piper method. The results were compared with two samples of Oxford Clay derived soils (Table II). The location of the sample sites is shown in Fig. 1. The results show a close similarity between the size range of Otmoor and Oxford Clay soils, allowing for the differences

TABLE II

		Per cent sample	
Otmoor			
1. HORTON (42/579142)			
Greater than	0.104 mm.	0.53	
	0.104 mm.–0.053 mm.	0.33	
	0.053 mm.–0.020 mm.	5.53	
	0.020 mm.–0.002 mm.	4.49	
Less than	0.002 mm.	85.94	Vegetable matter 3.18%
2. BECKLEY (42/573131)			
Greater than	0.25 mm.	1.00	This sample may be slightly biased by its proximity to Roman Road
	0.25 mm.–0.104 mm.	2.00	
	0.104 mm.–0.053 mm.	0.80	
	0.053 mm.–0.020 mm.	3.50	
	0.020 mm.–0.002 mm.	2.75	
Less than	0.002 mm.	80.00	Vegetable matter 9.75%
Marsh Gibbon Basin			
3. HEET FARM (42/652212)			
Greater than	0.25 mm.	4.50	
	0.25 mm.–0.104 mm.	3.70	
	0.104 mm.–0.053 mm.	5.60	
	0.053 mm.–0.020 mm.	18.65	
	0.020 mm.–0.002 mm.	20.32	
Less than	0.002 mm.	47.15	Vegetable matter 10%
4. CUTTERS BRIDGE (42/634215)			
Greater than	0.25 mm.	13.7	
	0.25 mm.–0.104 mm.	11.6	
	0.104 mm.–0.053 mm.	7.5	
	0.053 mm.–0.020 mm.	1.7	
	0.020 mm.–0.002 mm.	1.2	
Less than	0.002 mm.	64.6	

in vegetable matter content. They also show the differences between the composition of the soils in the upper and lower Ray basins. The clay components in all samples were built up similarly.

6. CONCLUSIONS

Otmoor has been considered an alluvial basin. The results given above do not confirm this conclusion. They show that Otmoor soils are more akin to Oxford Clay than those derived from alluvial matter. If it were an alluvial basin it would be reasonable to expect a higher proportion of coarse material. Otmoor appears to be part of the Oxford Clay lowland which has developed, under periglacial and fluvial conditions, as a normal clay plain. It had its own fan-like system of drainage until the nineteenth century. This system appears to have linked with the earlier Ray between the Oddington gap and the Islip gorge. The actual level of Otmoor may have been slightly affected by earth movements along the Wheatley fault which have gone on uring recent historic times (e.g. Wallis, 1665–6, *Phil. Trans.* No. 10, . 166).

Any ponding back of the Ray drainage by Cherwell flood-waters would e most likely to affect the river immediately above the gorge at Islip. Here es the expanse of Gallows Brook gravel (Sandford, 1926, 142) and the atch of 'Ray gravel' (Arkell, 1944, 63). The Ray terrace and the assoiated alluvium lying to the north of the periclines appear to indicate that tmoor was probably never a lake basin, or similar deposits could be xpected on it.

It is probable that Otmoor was never more of a marsh than other parts f the Oxford Clay lowland. The present backward state is the result of oor drainage by an incomplete system of dykes. Young was right in hinking that if it were properly drained it would be more productive. arts of the moor are cropped for hay but much of it is in rough grazing.

ACKNOWLEDGMENTS

We would like to thank all those who have assisted our work, in particuar the local farmers and landowners, also Prof. E. W. Gilbert, Dr. K. S. andford, Dr. M. M. Sweeting, Mr. D. F. Baden-Powell, Mr. R. J. horley, Mr. P. H. T. Beckett and Mr. J. C. Almgill.

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The Petrology of the Oil-Shale Group Sandstones of West Lothian and Southern Fifeshire

by J. T. GREENSMITH

Received December 1959

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ABSTRACT: The sandstones of the Scottish Oil-Shale Group are shallow water, deltaic-like ortho-quartzites with sub-ordinate subgreywackes. There is a gradation between these sandstones and limestones. The mineralogy suggests a source area consisting of sediments and lavas possibly of Lower Old Red Sandstone age. The flow pattern derived from the cross-bedding indicates that the material was mainly transported from a north-easterly direction and laid down on uneven bottom slopes. Slump structures, ripple marks, worm tubes and load casts are common. The lithification of he sandstones appears to have been relatively rapid and the pattern of cementation is discussed in detail.

1. INTRODUCTION

THE OIL-SHALE GROUP of the Midland Valley of Scotland is a facies of the Carboniferous Calciferous Sandstone Series and attains its greatest thickness in the Lothians area adjacent to Edinburgh. Approximately 5000 feet of sandstones, oil-shales, shales, marls and limestones are present and have been the subject of investigation for over one hundred years. With rare exceptions the studies have been stratigraphical, palaeontological and structural and the only detailed petrological papers are on the Houston Marl (Teall, 1925) the limestones (Greensmith, 1960) and a sandstone dyke near Queensferry (Greensmith, 1957). Young (1867) and Craig (1888–93) have discussed briefly the chemical and physical properties and Bosworth (1913) and Mackie (1923) have recorded heavy minerals from some of the sandstones. Moreover, Bosworth (ibid., 61) suggested a north-easterly source for the clastic material and this inference has now passed into most subsequent palaeogeographical papers (Macgregor, 1931; Richey, 1935; George, 1958, 1960; and Kennedy, 1958). This new petrological study confirms the validity of this concept.

Sandstones form between 30 and 40% of the mixed succession of rocks and are only a little less abundant than the varieties of shale. They occur not only in thick and widespread wedges, such as the Binny Sandstone, but also form thinner and more localised intercalations at various horizons within the succession. Marls and limestones are much less significant volumetrically.

The present work is concerned with regions on the north and south shores of the Firth of Forth. The sandstones were examined from a belt one to two miles wide extending from Kinghorn in the east to Rosyth in the west on the northern side, and on the southern side from a four miles wide belt extending from Granton, Edinburgh, westwards to Blackness.

Fig. 1 indicates the general stratigraphical succession and horizons from which the sandstone samples were taken. The precise placing of the samples within the upper part of the Oil-Shale Group created no great difficulty because of the relatively detailed structural knowledge of the productive shale fields (Carruthers, 1936; Kennedy, 1943), but less precision was possible in the lower part below the Burdiehouse Limestone particularly between the Long Craig Sandstone and Granton Sandstone.

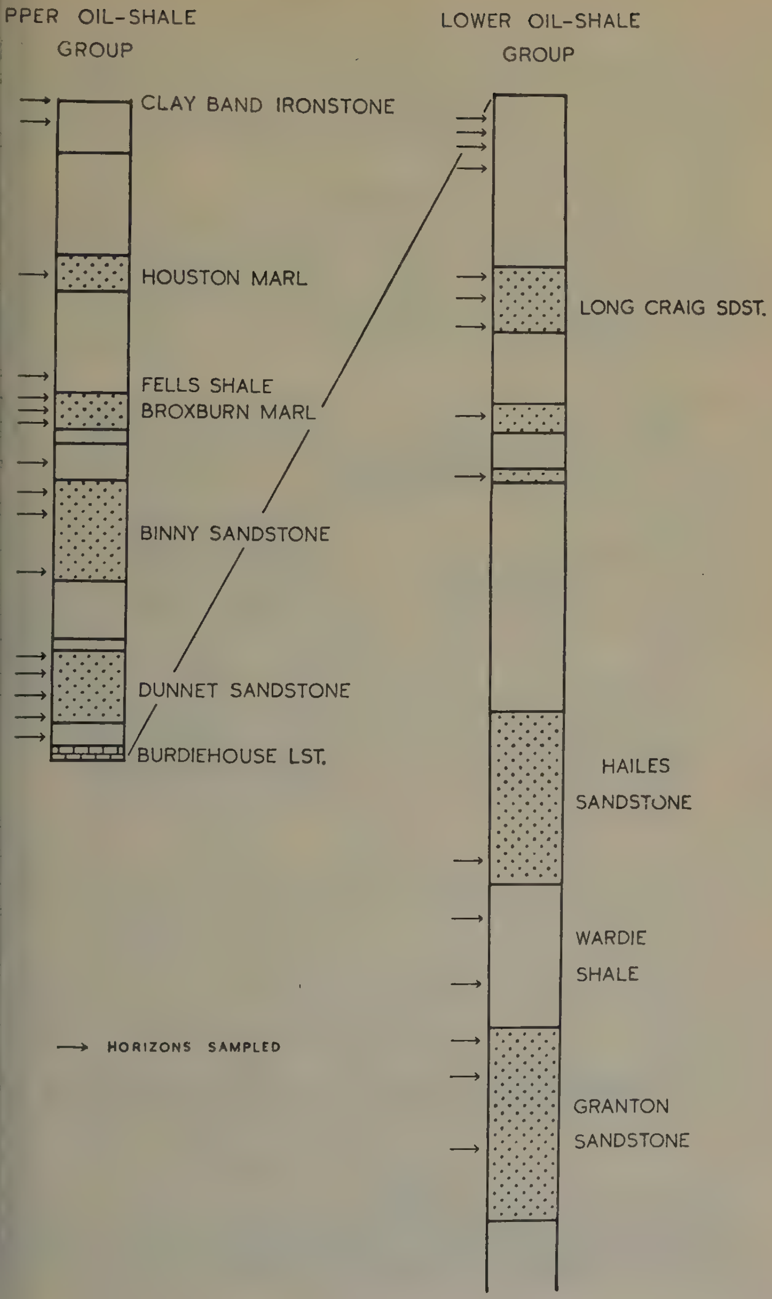
2. PETROLOGY OF THE SANDSTONES

(a) General Lithology

The most common characteristics exhibited by the sandstones in the field are lenticular and cross-bedding together with ripple marks. All these suggest an input of material transported by and deposited in relatively shallow waters. Furthermore, there is considerable lateral variability in the thickness of the beds and the Binny Sandstone, for instance, thins from 360 feet at Philpstoun[1] (049770) to 23 feet when traced for one and a half miles in a westerly direction (Kennedy, 1943, 14–15).

The textural nature of the sandstones as determined in thin sections is in keeping with the suggested shallow water conditions and most are well sorted. Fig. 2 illustrates by a representative series of cumulative curves the range in median diameter (0.068 mm. to 0.232 mm.) and shows the more significant arithmetic quartile deviations (QD_a) which are an expression of the degree of sorting. All QD_a values below the arbitrary figure of 0.100 mm. indicate good sorting and in none of the beds examined has 0.083 been exceeded. The best sorting is found in some layers of the Long Craig Sandstone at Long Craig Pier (144787) with values of 0.023 mm.

[1] National Grid references lie within the 100 km. square N36 (NT).



ig. 1. Generalised stratigraphy of the Oil-Shale Group area and approximate levels f sampling. (Vertical scale, approximately 1 in. to 250 ft.)

Naturally, these statistical data cover only the bulk of the beds as there are gradations at most localities into siltstones and at a few localities into conglomeratic layers which change the median grain size and affect the degree of sorting.

(b) Conglomeratic layers

Conglomeratic layers are uncommon within the area studied and are usually confined to the basal layers of a particular sandstone lenticle. At The Warrens (162791) to the south-east of Hound Point rounded pebbles of subjacent sandstones are present with a diameter up to 15 mm. These are a clear indication of penecontemporaneous erosion and deposition. Of

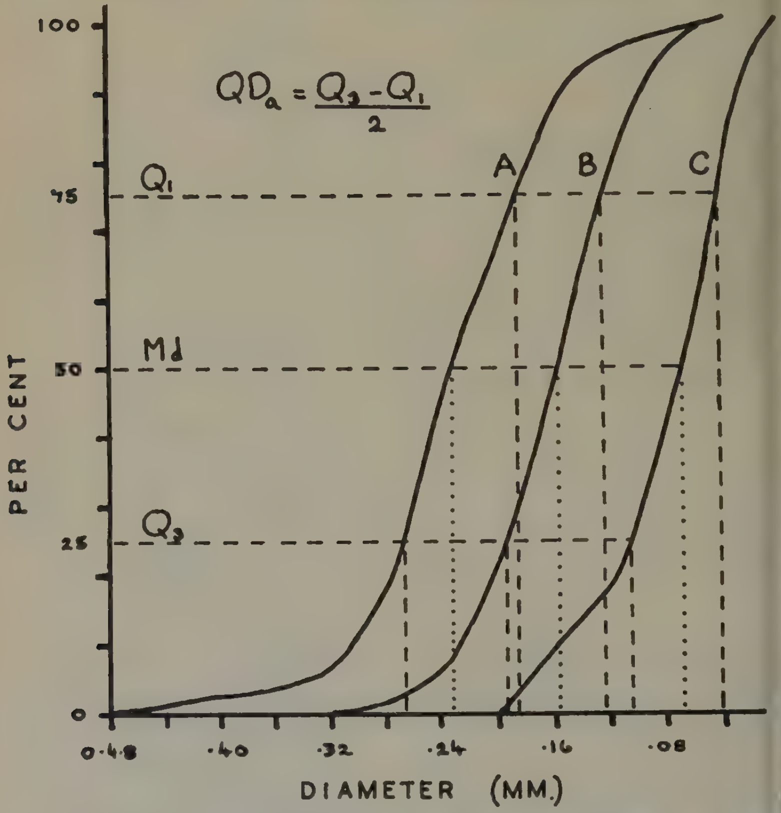
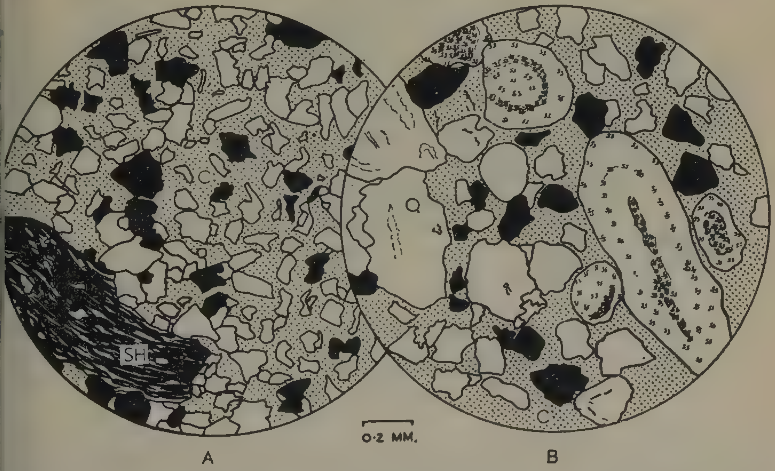


Fig. 2. Cumulative curves illustrating grain size and sorting of typical sandstones A: sandstone at The Warrens, middle of the Lower Oil-Shale Group; Md 0.232 mm. QD_a 0.040 mm. B: Dunnet Sandstone, Society; Md 0.154 mm., QD_a 0.036 mm C: sandstone at Blackness, Clay Band Ironstone horizon; Md 0.068 mm., QD 0.031 mm.

imilar origin are smaller shale pebbles within the Dunnet Sandstone at ociety (098791) (Fig. 3) and the Granton Sandstone at Craigleith (227745).

Perhaps the most peculiar penecontemporaneous conglomerate crops ut on the foreshore at Midhope (082795) towards the base of the Broxburn Marl horizon. It has a distinctive mottled brown appearance which on nspection proves to be due to the incorporation of strongly limonitised arbonate pebbles, the largest noted being 5 mm. in diameter, within a fine lastic matrix. The pebbles vary in shape from spindly to ovoidal (Fig. 3) nd sometimes exhibit a vague concentric structure internally which uggests that they may be large ooliths. Many show distinct evidence of vear which probably indicates derivation from some underlying bed.



ig. 3. A: shale pebble in the Dunnet Sandstone, Society. Crossed nicols. B: oolitic onglomerate, Broxburn Marl, Midhope. Crossed nicols. (Q: quartz, C: calcite, C: limonitised carbonate, SH: shale)

In contrast to the above varieties of conglomerate there are exposed in he Harwood Water section at West Calder (030638) conglomerates the naterial of which is of entirely different origin. They occur within the inny and Dunnet Sandstones and are characterised by rounded pebbles up to 16 mm. diameter) of weathered andesite and basalt, granophyre, uartzite and silty shale (Fig. 4). Their composition and generally rounded hape imply derivation from further afield.

(c) Sandstones

Mineralogically the sandstones throughout the Oil-Shale Group are elatively uniform, the only variation being in the amounts of carbonaceous natter and cementing agents. The source of the material, therefore, seems to have remained constant in character during the period of sedimentation.

Quartz is the dominant non-organic clastic component and varies in percentage from 34 to 99. At the higher values, and taking into account the good sorting, there is little doubt that the sandstones fall into the ortho-quartzitic grouping of Pettijohn (1949, 237) but towards the lower values the classification is more difficult. If the cements are ignored the sandstones remain ortho-quartzites in the strictest sense; if not, the weakly quartz-bearing with dominant cement, particularly where that cement is carbonate, could be termed sandy limestones (or dolomites) with equal validity.

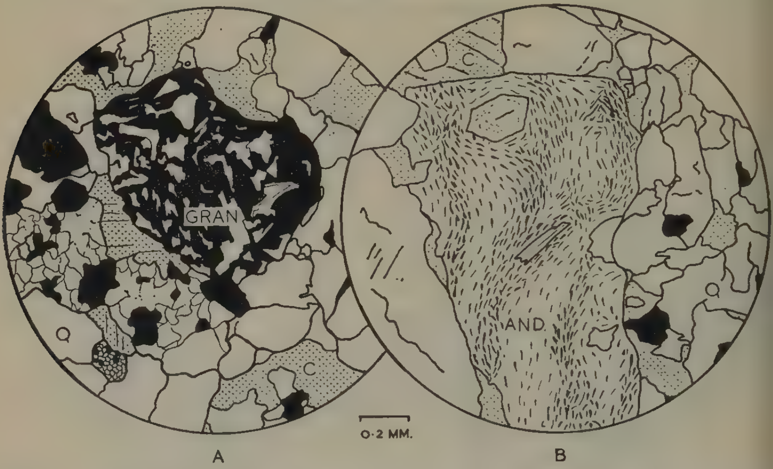
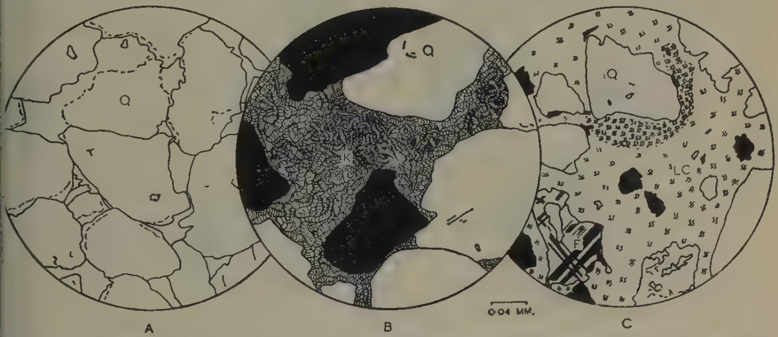


Fig. 4. A: granophyre pebble, Binny Sandstone, West Calder. Crossed nicols. B: flow banded andesite pebble with chloritised euhedral ? amphibole phenocryst, Binny Sandstone, West Calder. Crossed nicols

The Dunnet Sandstone at Grange Quarry, Burntisland (223867), excellently exemplifies the ortho-quartzitic texture which is so common in many of the other sandstones (Fig. 5 A). This texture is usually characterised by closely packed clastic quartz grains which are firmly welded by mutual intergrowth and authigenic quartz in the form of intermittent veneers. Slight modifications due to other cements may occur; an example of this is seen at the Clay Band Ironstone horizon near Blackness (049802). Here the typical ortho-quartzitic texture is modified by the presence of small pockets of a kaolinite-like mineral (Fig. 5 B). The same textural modification is found in pale brown, highly quartzose ganister layers which are sometimes present within or at the top of a sandstone. Though these ganisters imply conditions appropriate for the growth of vegetation, overlying coal seams are usually absent.

In complete contrast to the ortho-quartzitic sandstones are the open extured, carbonate rich sandstone layers exemplified by the Broxburn Marl horizon at Abercorn (082795) and below the Burdiehouse Limestone t Hopetoun (093792). In the former the clastic quartz and feldspar grains enerally 'float' in a matrix of partially limonitised fine-grained carbonate Fig. 5 C). A few of these clastic grains show a darker brown limonitised, oncentrically zoned veneer up to 0.20 mm. wide. This veneer is interpreted s a modification of the normal oolith structure found in many of the limetones (Greensmith, 1960), thus further emphasising the passage between he sandstones and the limestones of the group.



ig. 5. A: ortho-quartzitic texture, Dunnet Sandstone, Burntisland. Plane polarised ight. B: kaolinite pockets (K) in sandstone at Clay Band Ironstone horizon, Blackness. rossed nicols. C: 'floating' grains of quartz and feldspar (F) in carbonate matrix, roxburn Marl, Abercorn. Crossed nicols

However, textural variation in the sandstones is not confined to the arrow limits described above since opaque carbonaceous matter (up to %) often disrupts the simpler textures and when accompanied by clay inerals and carbonates creates ill-sorted varieties which tend towards ubgreywackes (Pettijohn, 1949, fig. 66). Representative of this type are the pper layers of the Granton Sandstone at Craigleith (Fig. 6 A), the top of he Broxburn Marl and Fells Shale sandstones at Midhope (Fig. 6 B), nd sandstones within the Wardie Shale west of Granton. These suggest ore rapid deposition than usual and the subsequent blanketing of the lastic material, allowing less time for winnowing agents to become ffective.

Despite the manifold textural changes resulting from localised modifiations in the depositional conditions it is possible to consider the mineralogy f all the sandstones in broad terms. For instance, in most of the beds the riginal clastic quartz grains are predominantly sub-angular to subounded, which suggests relatively rapid transportation and deposition with a low degree of abrasion. At coarser levels, where median grain sizes are greater than 0.20 mm., the originally deposited grains tend to be less angular, and any sporadic grains greater than 1.00 mm. are invariably well rounded. The presence of these larger, well-rounded grains, intermingled with the smaller sub-angular material, is not considered of any genetic significance as it is a feature which appears to be common to most sandstones in the geological column.

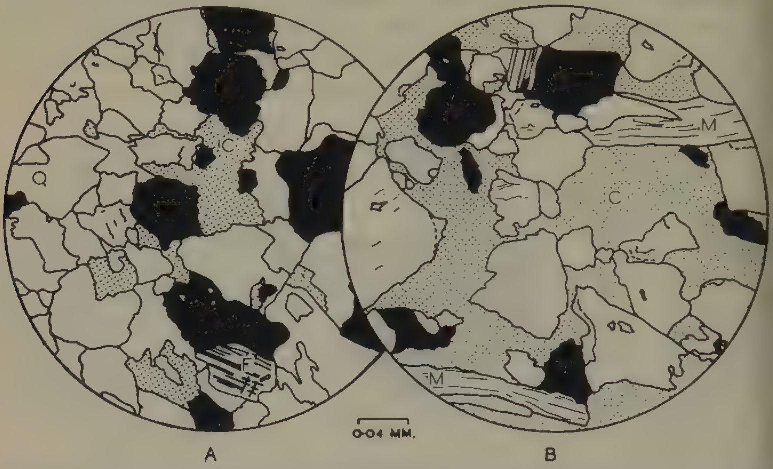


Fig. 6. Subgreywacke textural trend in A: Granton Sandstone at Craigleith, and B: Fells Shale sandstone at Midhope. Crossed nicols. (Q: quartz, F: feldspar, M: muscovite, C: calcite)

The outlines of many of the quartz grains have been altered by subsequent authigenic growth which may obscure their original shape; where detectable the new quartz is always in optical continuity. Because of this growth the shapes of the grains are extremely variable and the sphericity ranges from 0.2 to 0.9 (Krumbein & Sloss, 1951, 81).

As an aid towards determining the history of the grains a tentative threefold classification has been erected which may have some genetic significance. The grouping is as follows:

(i) unstrained quartz grains:	complete extinction under the microscope.
(ii) strained quartz grains:	incomplete extinction.
(iii) compound quartz grains:	welded aggregates, each constituent grain being optically dis-oriented from its neighbours.

All the grains fall into one or the other of these categories although the bulk are unstrained and the proportional distribution of quartz varieties remains roughly constant from top to bottom of the group.

Of course the genetic significance of this grouping is debatable yet it seems reasonable to suggest that the compound grains could have been derived originally from metamorphosed rocks. This is particularly certain where strong internal shearing is present. Though less definite, it is possible hat the unstrained grains came from granitoid rocks and the strained from variety of metamorphic rocks, bearing in mind, however, that some magmatic igneous rocks yield strained grains and some metamorphics unstrained. Nevertheless, on the basis of these generalisations it does seem probable that the grains were derived originally from a terrain composed predominantly of granitoid igneous and quartzose metamorphic rocks. This inference obviously ignores their subsequent history but its validity is enhanced by the presence of abundant apatite, rutile and sillimanite inclusions within many of the grains.

The feldspar content of the sandstones never rises above 4% and this can hardly be due to subsequent intra-stratal solution effects or weathering as he feldspars are usually fresh. In general the sandstones below the Burdiehouse Limestone are less rich in feldspar than those above, which may be a reflection of slight changes in the nature of the source area. On the other hand, such small changes may represent mere fortuity in sampling.

Angular, clastic microcline, orthoclase, oligoclase and andesine have been identified; in no instance is there any evidence of authigenic origin. Partial sericitisation and kaolinitisation are very rare and are usually confined to orthoclase. Thus, in all their characteristics the feldspars suggest a low proportion of strongly feldspathic rocks in the immediate source area. An alternative interpretation is that mechanical and chemical winnowing-out of the more unstable feldspar species could have occurred during weathering at the source or during transportation and deposition; but contradictory evidence is shown by the associated quartz grains which suggest a short period of abrasion insufficient to bring about wholesale feldspar breakdown.

Clastic mica flakes (muscovite, hydrobiotite and ? sericitic mica) are more common than feldspars in all the sandstones and frequently coat bedding planes. The muscovite in particular always appears unaltered and because of the inferred rapid transport and depositional conditions this can only mean that some of the source rocks were micaceous in character.

Finally, the heavy minerals require a brief mention even though the emphasis in this study has been on the main mineral components and their mutual relationships. The random samples analysed were taken from the full range of sandstone horizons and each was studied qualitatively, thus avoiding possible difficulties over authigenesis and intra-stratal solution which may have played some part in the final quantitative distribution of the heavy minerals. Consequently, a statistical evaluation of the minerals has not been attempted. Because of this the heavy minerals can be treated as a whole and, indeed, are easily dealt with in this fashion as they do not vary qualitatively throughout the succession. In all the sandstones tourmaline, zircon and topaz are present, and in the majority anatase, monazite, apatite and rutile may also be found. The Granton Sandstone at Craigleith, furthermore, yielded a colourless to pale brown staurolite.

With the exception of a little fracturing or chipping the grains are invariably sub-angular to well rounded. Very few have been detected which show the euhedral shape normally taken as suggesting either authigenic origin or derivation from a freshly exposed igneous or metamorphic terrain. Bosworth, however, has recorded angular tourmaline, rutile and zircon from these rocks at four localities (1913, 58–9), but despite his findings and resulting from this present wider examination the general implication is that the bulk of the heavy minerals have probably been derived from earlier sediments at least once in their history.

The most obvious and abundant heavy mineral in the separations, and in thin-sections, is a colourless to olive-green tourmaline. It is usually well rounded and ovoid with a maximum diameter ranging up to 0.45 mm.

Almost equally common are colourless to pale purple, sub-rounded to well-rounded zircons (up to 0.20 mm. diameter). Some are spheroidal whereas others are prismatic or tabular with worn bipyramids; both zoning and inclusions are rare.

It is the persistence of these two minerals which again suggests that the source area must have had a generally uniform petrological character.

(d) Cementation

The cementation of sandstones is a controversial subject even when the materials under discussion are relatively fresh. If the samples are from weathered outcrops the problems are enhanced and, as this applies to all the present examples, only tentative conclusions have been drawn about their relationships and origins.

The cements are numerous and can be treated in general terms simply because they are all present in every sandstone examined. Naturally, these cements are patchily distributed within any given sandstone and some layers, for instance, may be richer in carbonates or quartz to the exclusion of carbonaceous matter or clay minerals; the succeeding layer may show the reverse relationship, and so on. This type of cementation is, perhaps, seen at its best in the ortho-quartzitic and carbonate-rich sandstones such as the Dunnet Sandstone on the foreshore at Hopetoun (086795), whereas in the subgreywacke varieties it tends to be obscured, e.g. Granton Sandstone at Craigleith.

Taking the sandstones as a whole there is no single predominant cement and in each bed there is invariably an assemblage of both primary and secondary materials (Table I).

TABLE I. **Cements and order of cementation**

RIMARY	Carbonaceous and bituminous matter
	Iron-rich dolomite
	? Sericitic mica
	Kaolinite
	Chalcedony
ARLY DIAGENETIC	Authigenic quartz
	Authigenic dolomite
	Authigenic calcite
	Authigenic kaolinite
ATE DIAGENETIC	Authigenic limonite

Black, structureless carbonaceous matter is ever present and sometimes ggregated into distinct laminae; generally, however, it is disseminated in n irregular manner. Golden-brown to yellow bituminous fragments are nuch rarer and only appear in any quantity in the more silty layers of the andstones.

The most abundant primary cement is what appears to be an iron-rich lolomite ($\omega = 1.679$–1.683), which sometimes forms up to 40 or 50% of the ock. It usually occurs as a patchily distributed fine-grained, homogeneous ggregate and the individual grains are rarely greater than 0.01 mm. Less ommonly, when the percentage of carbonate is high, the distribution is ven and the clastic grains take on a 'floating' relationship to the matrix. In hese circumstances it is probable that there has been contemporaneous leposition of the clastics and carbonate and the latter probably represents chemical precipitation as there is no evidence of bioclastic origin.

Sericitic mica (? illite) is rare and its presence may result from the breaklown of feldspar or muscovite even though it sometimes abuts against erfectly fresh grains of those minerals. Orthoclase is the only feldspar vhich does show partial sericitisation but its post-depositional contribuion must be small as it is an uncommon mineral. Therefore, it seems likely hat most of the serecitic mica is primary in the depositional sense though ts origin remains obscure.

The kaolinite-like minerals present similar problems. The minerals as a group have been optically identified by their low birefringence, straight extinction and characteristic rouleaux texture. Sometimes these properties re obscured by a rather diffuse light brown stain which is possibly a veneer f secondary limonite. Even so it is clear that the mineral group has a widespread distribution in all of the sandstones. Invariably it takes the form of rregular, interstitial pockets, some of which measure 1.50 mm. by 0.30 mm. These pockets may abut against fresh feldspars and micas and this fact suggests two possible modes of origin. Either the kaolinite-like minerals were simultaneously precipitated with the clastics as primary gels and have subsequently consolidated or they were precipitated later in available pores as a result of migration in solution from some external or internal source (Hemingway & Brindley, 1948). Of these two hypotheses the latter is favoured simply because of the absence of kaolinitised feldspars and the fact that the minerals often occupy the centres of cavities or pores which are already lined with early diagenetic minerals such as quartz and dolomite (Fig. 7A). Additionally, it is now well established that the presence of abundant calcium ions in solution, as there undoubtedly were during Oil-Shale Group times, inhibits the formation of kaolinite minerals (Millot, 1942). Therefore it is assumed that they were subsequently emplaced in the sandstones from some source which remains obscure.

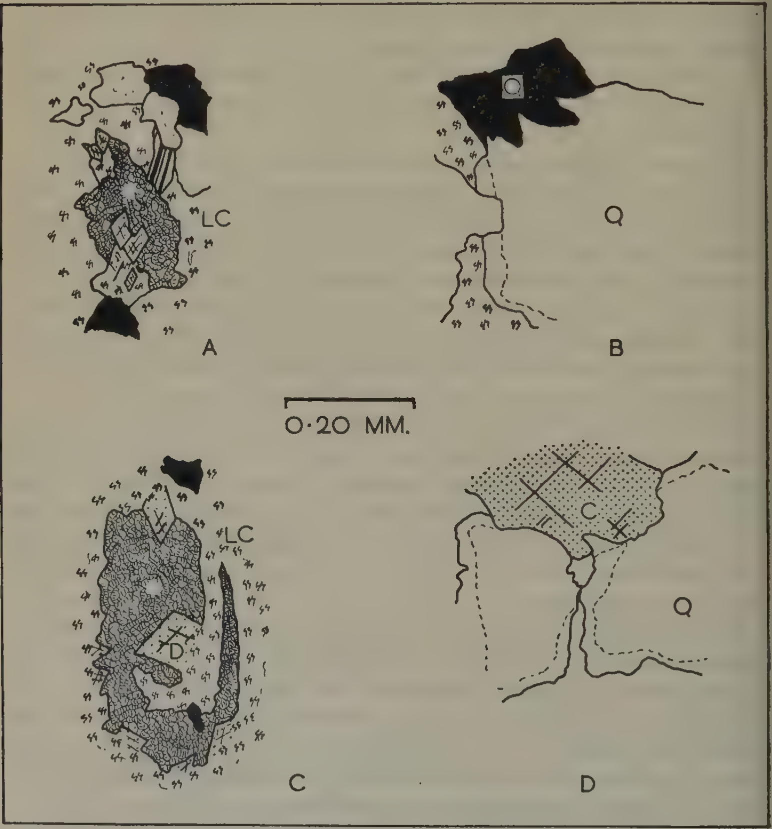


Fig. 7. A: kaolinite infilled pocket, Queensferry Cement horizon, South Queensferry. Crossed nicols. B: quartz interpenetration, Broxburn Marl horizon, Abercorn. Crossed nicols. C: kaolinite infilled pocket with marginal authigenic dolomite rhombs (D) grading into limonitised carbonate (LC). Sandstone below Burdiehouse Limestone, Hopetoun. Crossed nicols. D: calcite (C) replacement of clastic and authigenic quartz. Sandstone below Burdiehouse Limestone at Hopetoun. Crossed nicols

Chalcedonic silica is a less common cement than clay minerals and generally occurs in pockets smaller than 0.20 mm. by 0.15 mm. Characteristically these pockets are ovoid which suggests that they were deposited before compaction of the sediment. There appears to be no marginal replacement by carbonate or, conversely, replacement of the matrix by chalcedonic silica so that it is possible that the mode of origin was primary precipitation in gel-like form. In the early phases of diagenesis progressive solidification of the gel occurred and terminated after the deposition of authigenic quartz.

Authigenic quartz in the form of outgrowths on the clastic quartz grains ranks equally with the primary carbonates as a cementing agent (Fig. 5 A). The new growth is always in optical continuity and it leads, in many sandstones, to a high degree of grain interlocking and the development of good quartz-mosaic or ortho-quartzitic texture. The crushing and contortion of clastic mica flakes and more particularly the development of good euhedral pyramidal terminations by some of the grains indicates that growth took place soon after clastic deposition whilst adequate free pore space was still available.

Despite this widespread effect it is not unusual to find the original clastic grains in contact with each other although there seems to be little interpenetration and practically no suturing (Fig. 7 B). Because of this it is unlikely that all the new quartz has developed as a result of pressure solution of the clastic quartz followed by reprecipitation in adjacent pores (Waldschmidt, 1941; Greensmith, 1957; Thompson, 1959). Perhaps direct precipitation from silica-rich waters may account for the greater part of the new growth in the manner suggested by Bien, Contois & Thomas (1959). Their preliminary conclusions are that there are two probable mechanisms of inorganic precipitation. One involves the reaction of soluble silica with electrolytes in sea-water and the other adsorption or co-precipitation of soluble silica with suspended solids or colloidal matter in river-water again as a result of contact with electrolytes. Either of the two hypotheses may be appropriate for the Scottish sandstones as marine waters were probably never far away from the area of their deposition. Certainly they appear to approximate closer to the then prevailing circumstances than Siever's suggestion (1957) of bio-chemical quartz precipitation by means of organisms (radiolaria, sponges, etc.) which concentrate amorphous silica in their tests. He visualised that after the death of the organisms and consequent deposition of their tests a recrystallisation of the amorphous silica occurred which is commonly expressed in the form of quartz outgrowths. As there are no traces of siliceous organisms within the sandstones this particular origin seems unlikely.

As the proportion of primary carbonate cement increases the proportion of authigenic quartz decreases and in the sandstones with more than 30%

of carbonate new quartz is rare. When present there is no doubt that it was formed after the primary carbonate because of the 'trapping' of carbonate grains between the core and the new veneer.

Authigenic dolomite is not common and always post-dates authigenic quartz precipitation. In most cases the large euhedral and subhedral rhombic dolomite grains (up to 0.20 mm. maximum diameter) appear to be derived from the solution and partial reprecipitation of the primary carbonates as there is a physical gradation between the two (Fig. 7 C).

On the other hand, authigenic calcite has a more dubious origin and there is no gradation into primary carbonates. Possibly its source is calcareous shell fragments but none have been observed within the sandstones. This would suggest that direct chemical precipitation in pores from infiltrating calcium-rich solutions is more probable. The calcite is normally coarse grained, anhedral and, apart from occurring in irregular interstitial patches, is found in tenuous veins. In contrast to all the other carbonates it invariably appears to corrode feldspars and quartz although chalcedonic silica pockets remain unaffected (Fig. 7 D).

The actual dating of the secondary calcite cementation cannot be successfully demonstrated in the sandstones and, with the exception of the authigenic quartz, the same may be said about the other secondary cements. The new quartz is probably early diagenetic but the emplacement of the others could have taken place soon after the sediment had been deposited (early-) or at any time subsequently (late-diagenetic).

The limonitisation processes which have affected the iron-rich dolomite and hydro-biotite are similarly difficult to date but are almost certainly recent. The changes are probably a result of the lowering of the water-table and an incoming of oxidising conditions (K. C. Dunham, 1955, personal communication).

One of the apparent anomalies of the sandstones in view of the range of cements is the general absence of the montmorillonitic group of clay minerals. This is unexpected because igneous activity was relatively commonplace within and adjacent to the depositional area. Only at the Clay Band Ironstone horizon west of Blackness does a possible green and fibrous montmorillonitic mineral appear and there it merely has a localised and patchy distribution. It may be of some significance that this horizon is near the top of the Oil-Shale Group since contemporaneous intermediate and basic lavas and tuffs were being ejected in the Burntisland area, fifteen or so miles to the east-north-east.

(e) Sedimentary Structures

(*i*) *Lenticular and Cross-bedding*. Structural features indicative of relatively shallow water deposition are common within the sandstones; of these, lenticular and cross-bedding are the most obvious in the field.

The lenses invariably exhibit erosion surfaces at the top and the base and ometimes fill recognisable wash-outs. This is well seen at the type-locality f the Long Craig Sandstone to the east of the Forth Bridge and Page (1865) so recorded one at Hailes Quarry (208703) in Edinburgh.

The thickness of each individual lenticular unit varies up to five feet but ocally a thickness of about ten feet is attained as, for example, at Craig-ith Quarry, Edinburgh. Similarly, the normal maximum diameter or long xis is twenty to forty feet with local maxima of 150 to 180 feet. In all cases e direction of elongation is approximately parallel to the dip of the ternal cross-bedding if this structure happens to be present. It is usually ell displayed within each unit and the fore-set beds are generally linear r slightly concave upwards. At Craigleith, Society, South Queensferry 32784) and Dalgety Bay (166837) a stronger concavity is present within ome of the beds and this is developed even further at Abercorn Point, orth of Hopetoun House, where the six-inch-thick lenses are distinctively stoon-bedded.

Wherever these structures can be observed in three dimensions the angles f fore-set dip and direction of the fore-set beds have been measured. One easurement was made for each lenticular unit displayed and, in all, it roved possible to make 103 readings from thirty-one exposures. Only one ccurate three-dimensional reading was possible at some localities whereas t others seven to eleven could be made.

The results of this preliminary work are of considerable interest in stablishing the broad areal flow pattern of the transporting waters. The orrected dips of the fore-set planes give values varying between 5 and 15° ut these may well have been higher at the time of deposition, prior to ompaction. However, they are of little significance when compared with e directional readings at each locality and on which the areal flow pattern actually based. Most of this directional information is plotted in Fig. 8 nd each arrow represents the average direction at that locality. The general istribution and pattern of the transporting currents becomes immediately ear and suggests that material was being swept in from a source to the orth-north-east or north-east. The regional slope of the depositional area ust have been from the north-east to lower levels in the south-west, at ast within the area studied. The anomalous northerly flow directions corded at Craigleith and Dalmeny (148774) may, on the other hand, dicate a localised reversal of that regional slope with material of southerly rigin, yet the petrography of the sandstones at those localities is no differ-nt from that of all the other sandstones. It is possible therefore that these versals merely express irregularities on the floor of the basin.

By plotting the azimuths of the fore-set beds of the sandstones of the wer and upper parts of the Oil-Shale Group separately it is also obvious at the broad flow pattern and thus, presumably, the source area remained

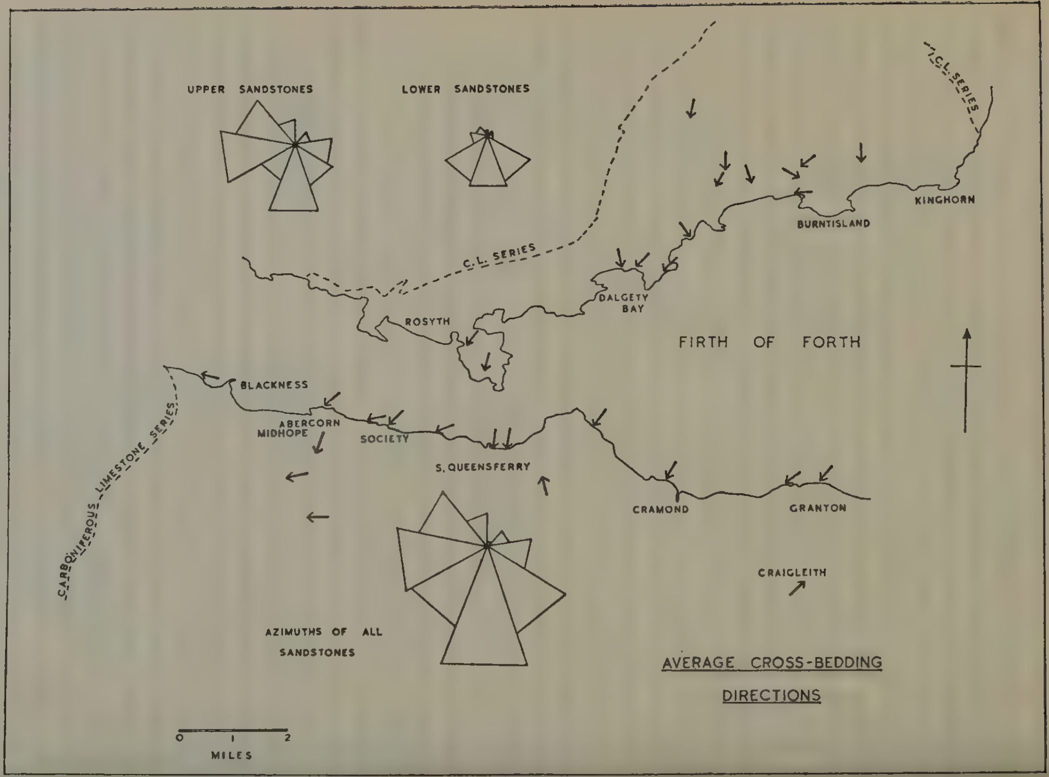
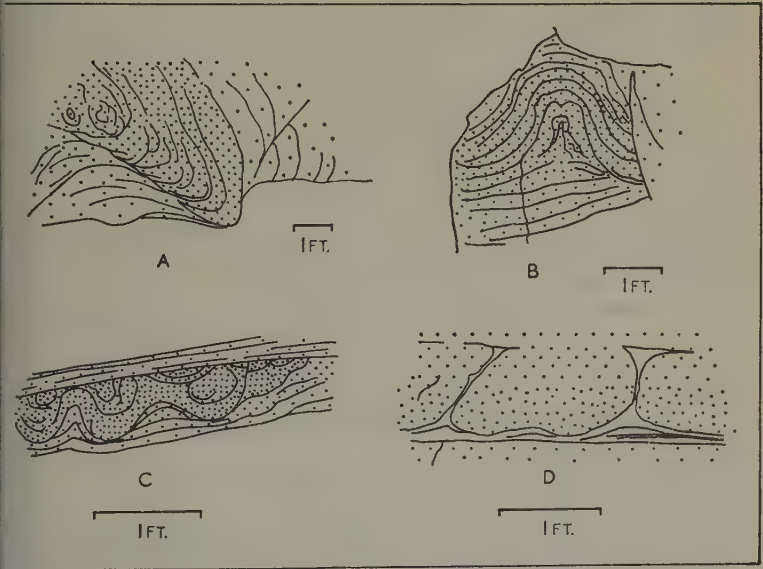


Fig. 8. Sedimentational flow pattern and azimuthal distribution of cross-bedding directions

nstant in position throughout most of Oil-Shale Group times (Fig. 8).
(*ii*) *Slumping and Flow Casting*. Numerically, the fore-set bed readings scussed above are low. This is not entirely due to a lack of three-mensional exposures but due, in many cases, to the restrictive influence of ump structures of variable magnitude which contort the bedding. This ntortion is well seen in a number of sandstones intermittently exposed on e shore from Granton harbour westwards to Eagle Rock, near Cramond 84775). Fig. 9 A and 9 B illustrate some of the usual expressions of is type of distortion and overturning which is by no means restricted to e lower stratigraphical horizons.



g. 9. Contorted slump bedding at A: Eagle Rock, Cramond, B: Sealscraig, South ueensferry, and C: north-east of Hopetoun House. D: load casts at Dalgety Bay

The slumping may be more precisely demonstrated in the form of flow sting (Shrock, 1948; Greensmith, 1956) as in a sandstone fifty or so feet elow the Burdiehouse Limestone to the east-north-east of Hopetoun ouse (095792). Within it and towards the base there are two distinctive vels eighteen inches apart with 'balled-up' or pillow-like bodies each of hich is up to eight to ten inches thick (Fig. 9 C). The sandstone laminae side the bodies are strongly contorted whereas externally the laminae of e parent sandstone are little disturbed apart from a slight squeezing pwards between the bodies. In each instance there is an overlying sand-one layer which is totally undisturbed and which rests on an erosion surface truncating the upper surfaces of the structures. Therefore, it is highly probable that these flow casts represent a penecontemporaneous, *en masse* movement of partially lithified material down shallow slopes on the irregular bottom of the depositional basin. It is a movement which may have been brought about merely by unstable loading on the bottom slopes or by localised thixotropic mobilisation during contemporary earth tremors (Boswell, 1949); either origin is feasible. Moreover, it is possible that most of the other slumped sandstones resulted from similar causes even though they do not exhibit obvious internal erosion surfaces.

(*iii*) *Load Casting*. Similar in external and internal form to flow casts but differing in the absence of a truncated upper surface are load casts (Shrock, 1948). They tend to be present in the area studied wherever sandstones overlie thin shales and are an expression of uneven loading during compaction, the coarser grained sandy material being squeezed into the underlying, probably more plastic, muds.

The best probable examples can be seen at the foot of the cliff below the old church at Dalgety Bay (170838) occurring at the base of a brown sandstone sixteen inches to two feet thick (Fig. 9 D).

The sandstone is continuously exposed for 150 feet and rests on four to five carbonaceous, shaly bands (a quarter to half an inch thick). It is these shaly bands which appear to have taken most of the compactive stress during the loading movements thus leading to distortion and eventual injection between adjacent casts. The casts are closely packed and much larger than those seen elsewhere, being anything up to two feet long by eighteen inches thick in two dimensions. At the eastern end of the cliff section a further intercalated shale lamina appears and this gives rise to another layer of casts.

(*iv*) *Ripple Marks*. Ripple marks of shallow water origin are a common feature of the sandstones and are particularly well developed within the Dunnet Sandstone at Port Neuk (138784), Banks (104788) and Society, the Binny Sandstone at Society, and the Broxburn Marl at Abercorn Point. Invariably the trend of the ripple crests is linear and they usually have a rough asymmetric cross-section. The ripple index approximates to six with wave-lengths of about three inches and amplitudes of about half an inch. This index value is characteristic of littoral-cum-neritic depositional conditions.

The direction of crestal linearity is closely related to the cross-bedding of the sandstones and is always at right angles to the dip of the fore-set beds, providing the latter have not been subsequently disturbed: as the fore-set beds vary in direction from bed to bed so do the ripples within any given outcrop. There is little doubt from these features alone that the bottom currents must have been fluctuating rapidly in strength and direction while the sand was being laid down.

(*v*) *Worm Tubes*. Tubes due to the boring activities of organisms are oorly preserved at outcrop and are only developed at thinly bedded, ighly carbonaceous and fine-grained levels within the sandstones. They e present in the beds at Long Craig Pier, at the base of the upper part of e Oil-Shale Group in Dalgety Bay and just below the Houston Coal orizon in Craigton railway-cutting (080763). As this type of lithological riation is commonplace it is probable that more borings will be disovered.

In Craigton cutting the tubes are abundant, relatively closely packed and enerally elongated in a near vertical position with a length varying between n and twenty millimetres. Their width is more or less uniform throughout e length and the circular to oval cross-section measures three to five illimetres. In each instance the infill of the tube is of non-carbonaceous nd and this may indicate a purifying action on the part of the organism.

With these characteristics, and taking into account their distribution in ner grained and more carbonaceous layers, it is probable that the borings e a result of the activities of worms. The worms possibly migrated into e area because of the quieter conditions then prevailing as compared ith the usual turbulent conditions. Similar tubes have been described in etail from Upper Carboniferous sandstones in northern Derbyshire Greensmith, 1956, 352–4) and there they can be ascribed to the organism *lanolites* (S. Simpson, 1956, personal communication).

3. PETROGENETIC SUMMARY

(a) Source and Transportation of Material

In the determination of the provenance of the clastic Oil-Shale Group ediments the most fruitful source of information are the sandstones, as by eir very nature they are derived sediments.

The overall picture presented by these sandstones suggests an input of aterial from a land area of relatively constant lithology throughout the eriod. This conclusion is favoured by the persistence of heavy mineral ecies such as tourmaline, zircon, rutile, anatase, topaz and monazite, gether with the constancy in the nature of the feldspars (microcline, rthoclase, oligoclase and andesine). Further confirmation comes via the istribution of the quartz grain varieties (strained, unstrained and comound) which remains more or less uniform from the bottom to the top of e succession.

Quartz is the most important constituent of the rocks and varies in ercentage between 34 and 99. The original clastic grains invariably seem be sub-angular to sub-rounded and usually have irregular shapes. Micas muscovite, ? sericitic mica and hydro-biotite) are the next most important onstituent, and muscovite, in particular, is often abundant. The feldspars, which never amount to more than 4% at any locality, are usually sub-angular and fresh.

Therefore, the implication is of source rocks with a highly quartzose, micaceous and slightly feldspathic nature, a condition which could be met by many rocks such as acid gneisses and granulites. On the other hand, the general lack of acute angularity in the quartz grains and the rounded nature of the bulk of the heavy minerals indicate a maturity of the sandstones which would hardly be attained without long continued or repeated erosion and transport. In view of these facts it seems much more feasible to suggest that the source rocks consisted predominantly of sediments such as sandstones and this conclusion has some confirmation in the presence of sandstone and siltstone pebbles in the conglomerates at West Calder. If the source area did have this fundamental nature it is equally likely that igneous rocks were also present to account for the freshness of the feldspars and the presence of degraded biotites. The granophyre, andesite and basalt pebbles in the Binny Sandstone at West Calder substantiate this viewpoint. The freshness of the feldspars is, moreover, an expression of the rapidity of erosion of those igneous rocks and subsequent rapid transportation and burial of the derived fragments.

The various shallow-water structures exhibited by the sandstones have a direct bearing on the composition and possible geographical position of the source rocks: there is no doubt that the mode of transportation was by water. Of these structures the cross-bedding is of greatest significance as it indicates that the regional basinal slope was approximately from the north-east to south-west. Most of the material presumably came into the area from the north-east although the localised reversal of the average cross-bedding at Craigleith and Dalmeny may be due to minor contributions from the south.

In fitting this inferential pattern of sedimentation with a dominant north-easterly source into its proper environmental framework a lack of factual evidence prevents any dogmatic conclusions. The detailed petrology of contemporary sediments in east Fife and of the underlying Upper and Lower Old Red Sandstones is unknown. It is known, however, that at least 19,000 feet of coarse, iron-rich arenaceous sediments were deposited under fluviatile and lacustrine conditions at the north-eastern end of the Midland Valley during Lower Old Red Sandstone times (Kennedy, 1958, 116). Interbedded with these sediments are andesites, sub-ordinate rhyolites and pyroclastic rocks of similar composition.

It is these rocks which may, in fact, be the source of the Oil-Shale Group sandstones. Compression of the Midland Valley and the development of a graben with parallel-trending, north-easterly Caledonian ridges and depressions occurred towards the end of Lower Old Red Sandstone times. At the same time uplift and erosion commenced to the north of the High-

and Boundary Fault of the cover of Lower Old Red Sandstone draped ver the crystalline basement rocks (George, 1960). The earliest sediments aid down in the depressions were red and brown conglomerates, sandstones, marls and nodular limestones. These are of Upper Old Red Sandstone age and were succeeded by the lowermost Carboniferous Cementstone Group shales, limestones and sandstones. At this stage, that is at the nd of Cementstone times and the beginning of Oil-Shale Group times, it s probable that the higher Caledonian ridges still projected above the evel of the accumulating sediments (Kennedy, 1958, 128; George, 1960, 7). It is tentatively suggested that one of these residual and rapidly decayng ridges may have been the source of most of the clastic material swept nto the depositional basin. Such a ridge lying to the north-east of the present area may have acted as a north-east to south-west barrier of ndeterminate length but so disposed as effectively to prevent the input of material from the older crystalline rocks now possibly exposed farther north. The actual position of the critical north-eastern extension of the ridge is quite conjectural but perhaps it now lies to the south of the presumed extension of the Highland Boundary Fault under the North Sea. The south-western end is more firmly fixed by George (1958, 301) who places the zero isopachyte for the Calciferous Sandstone Series between Lomondside and the Ochil Hills and south of the Ochil and Sidlaw Hills n Fife.

The postulated source-ridge satisfactorily accounts for the petrographic nature of the sandstones and, apart from their essential mineralogy, it explains the widespread distribution of calcium, magnesium and iron cementing materials. The abundant calcium and magnesium would be derived largely from the decay of the igneous lavas whereas the iron probably came from both the lavas and the range of iron minerals present in the Lower Old Red Sandstone sediments. Likewise, the genesis of the clay minerals such as kaolinite may be associated with the lava breakdown.

(b) Conditions of Deposition

The tectonic framework of the Oil-Shale Group sedimentation is relatively well known (Kennedy, 1948, 126). It involved a general downwarping of the whole area on which was superimposed localised flexuring due to renewed movement along Mid-Devonian fold axes.

On this unstable foundation shallow, warm lagoonal waters probably prevailed in which carbonates were being constantly precipitated to consolidate later as limestones (Greensmith, 1960). At intervals the pattern of chemical sedimentation was disturbed by deltaic-like inputs of clastic material mainly from the north-east. This material was laid down rapidly and unevenly by the transporting currents in deltaic fashion and the erosive power of these postulated currents was such that quite frequently wash-outs and shale conglomerates were formed. Sometimes the unstable weight of sandy material led to slumping down the shallow bottom slopes; at other times the weight was sufficient to contort the underlying beds. Often the sands built up to water-level and became capable of sustaining plant life.

The finer clastic material, exemplified by shales, probably represents quieter phases of deposition marginal to the more sandy parts of the distributaries. In fact, it has been suggested (L. R. Moore, personal communication) that the separation of the basin into lagoonal tracts may have been a direct result of this type of deltaic deposition, the sands acting as barriers between adjacent lagoons.

(c) Diagenesis

The main diagenetic effects observed in this study involve the problems of cementation which have already been adequately discussed. Generally, the lithification of the sandstones appears to have been rapid as suggested by the numerous penecontemporaneous conglomerates. This rapidity was not always matched in the shales which sometimes show the effects of upwards squeezing in load-cast structures so implying plasticity. Apart from this type of structure there is little direct evidence of compaction.

Intra-stratal solution changes have not been detected though this does not mean that they are absent. Local metasomatism due to subsequent basic intrusions is almost negligible and usually expressed by a secondary silicification of the sandstones marginal to the intrusions.

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The Mode of Emplacement of some Dykes and Veins in the Pre-Cambrian Rocks of Ceylon

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BSTRACT: Several dykes and veins of pegmatite and aplite occurring in the Pre-ambrian rocks of the Colombo, Anamaduwa, Galgamuwa, Wilgomuwa, Hembarawa nd Rakwana areas are described.

These dykes and veins show the following megascopic features:

(a) Absence of offsetting resulting from dilation at the intersections;
(b) Enclosure of undisturbed patches and blocks of country-rock in the dykes;
(c) Continuation of wall-rock structures (e.g. gneissic foliae, elongate minerals) cross dykes and veins;
(d) Truncation of basic patches and inclusions in the host rock and the absence or ack of correspondence of related portions within the dyke or on opposite wall;
(e) Marked variation of width, i.e. 'pinch-and-swell' structure;
(f) Coarse-grained margins and finer-grained cores;
(g) Sharp and diffuse contacts; reaction zones in wall rock.

These features, which occur singly or in combination, together with the microscopic letails, indicate that the dykes and veins have been emplaced non-dilationally and were ormed by replacement or static metasomatism. Some dykes may however show partial nobilisation.

The mode of emplacement of both 'primary' and 'secondary' ptygmatic veins from ome of these areas is also considered.

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1. INTRODUCTION

N THE COURSE of geological mapping in several scattered areas of Ceylon he writer has paid some attention to the dykes and veins of aplite and egmatite which intersect the country-rocks, and to their probable mode f emplacement. Many of the features exhibited by these dykes and veins losely resemble those observed in other Pre-Cambrian areas, e.g. the Osi rea of Nigeria (King, 1948), Cornucopia, Oregon (Goodspeed, 1940, 1952), and the Tananarive region of Madagascar (Laplaine, 1957).

It is generally recognised that such veins and dykes may be emplaced either by dilation, i.e. by magmatic injection along and widening of an initial fissure, or non-dilationally, i.e. by replacement. Goodspeed (1940), after a detailed study of many dykes ranging from diabase to pegmatite in the Pacific North-west, enumerated the structural and textural criteria by which either mode of emplacement may be recognised. These criteria are applied to examples drawn from several parts of Ceylon and the bulk of the evidence suggests that the dykes and veins described have been formed by replacement of the country-rocks and not by injection.

The dykes and veins are mainly pegmatites and aplites occurring in the granitic and granodioritic migmatites belonging to the Pre-Cambrian Vijayan Series. Examples are taken from the wide zone of migmatisation round the Tonigala Granite in the west, from the Colombo District, and from the Wilgomuwa–Sorabora–Hembarawa area in the east. A few examples of dykes and veins in the metasediments of the Khondalite Group of the Rangala and Rakwana areas are also given (Fig. 1).

2. MACROSCOPIC FEATURES

(a) Absence of offsetting

One of the most convincing arguments for a non-dilational mode of emplacement of dykes and veins put forward by Goodspeed (1940, 189, 190) is the absence of offsetting at the intersection of two or more dykes at an oblique angle.

This feature is seen over and over again in the Pre-Cambrian rocks of Ceylon and examples are given below from all the areas under consideration. At Potgulkande, near Hembarawa, for example, a number of narrow aplites occurring in quartzo-felspathic gneisses cross each other at very oblique angles without the slightest amount of offsetting at any of the intersections (Fig. 2A). Another feature is also well seen here, namely, the continuation of elongate leaves of quartz, some about a quarter of an inch long, from the gneiss across the veins, without interruption or distortion (Fig. 2A, inset).

Another example was observed at Nekettegama, near Tonigala (Fig. 2C), in which biotite granulite was crossed by a narrow aplite vein which in turn was intersected by a pink granite pegmatite containing recrystallised biotite and hornblende; no offsetting is evident. In a vertical section from a quarry at Kaduwela in the Colombo District (Fig. 3) the country-rock of biotite granite gneiss which contains concordant or nearly concordant veins and patches of pegmatite emplaced without distortion of the gneiss foliae is intruded first by a pink aplite and later a pink pegmatite. No offsetting takes place at any of the intersections, nor has vertical movement accompanied the emplacement of the aplitic dyke.

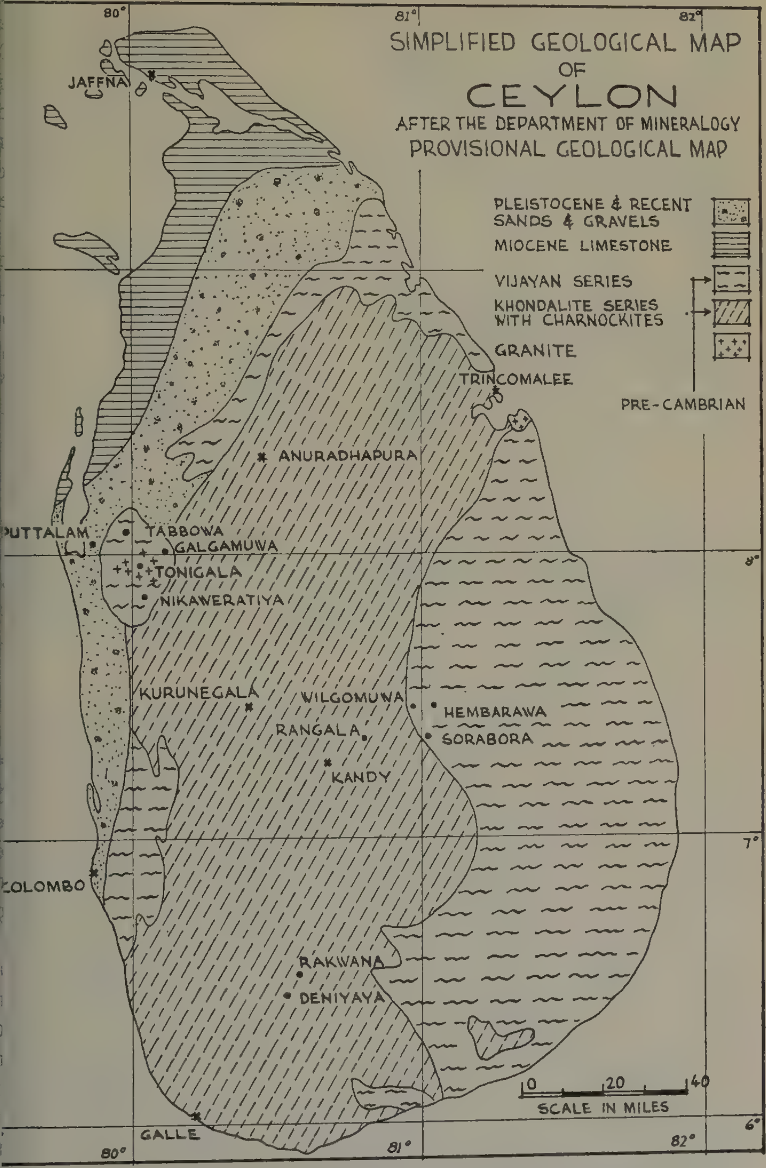


Fig. 1. Simplified geological map of Ceylon showing localities mentioned in text

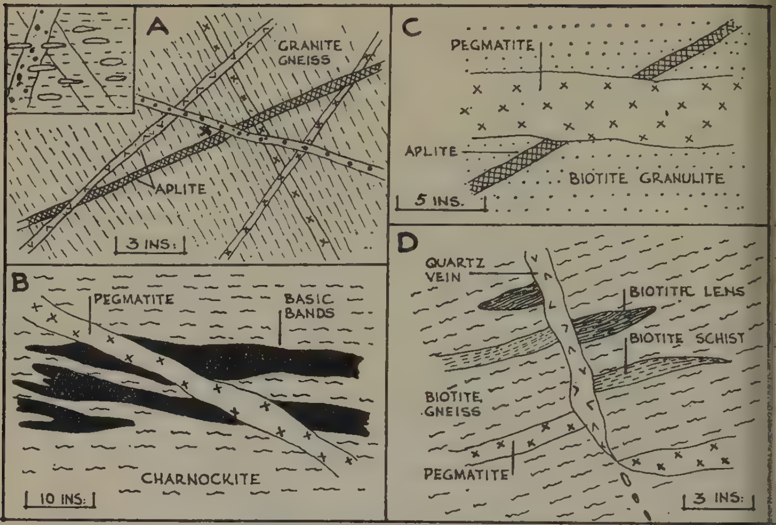


Fig. 2. A: Absence of offsetting at intersections of aplite veins. Inset: uninterrupted continuation of quartz leaves in gneiss across vein. Potgulkande, Hembarawa. B Pegmatite dyke crossing basic bands in charnockite without displacement. Milestone 90 on Colombo–Deniyaya road (vertical section). C: Non-dilational intersection o aplite and pegmatite. Nekettegama, Tonigala. D: Quartz vein emplaced along earl shear plane (vertical section)

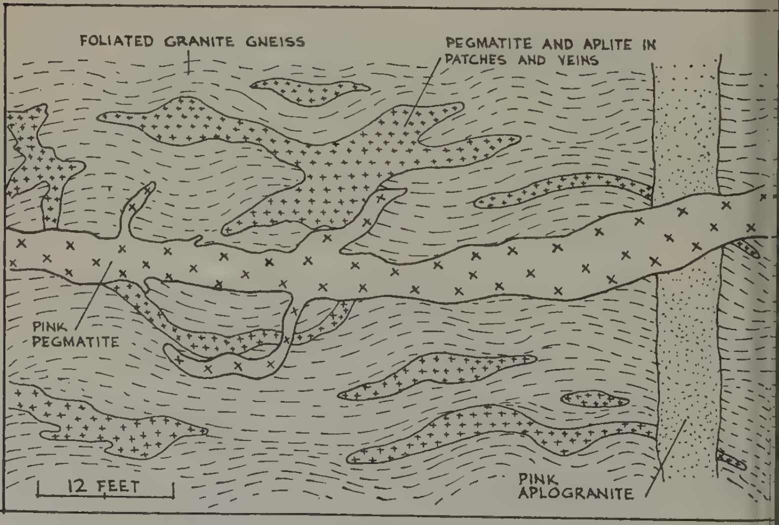
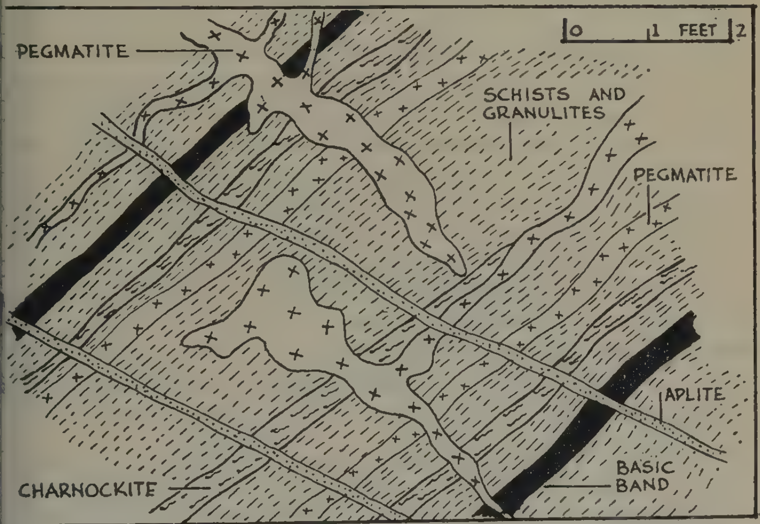


Fig. 3. Non-dilational emplacement of pegmatite and aplogranite in granite gneiss with early concordant pegmatite. Kaduwela, near Colombo. (Vertical section)

Examples from the metasedimentary Khondalite Group are also com- n. At milestone 90 on the road between Colombo and Deniyaya (Fig.), pegmatite dykes cross basic bands in charnockite at a very oblique gle without any offsetting. At a point 100½ miles along the same road, gmatites intrude Khondalite Group gneisses and schists non-dilationally g. 4). The phenomenon is also shown by the pyroxene–scapolite dykes the Rangala area where they cross each other or traverse basic bands in gneisses (Cooray, 1960).



g. 4. Non-dilational emplacement of pegmatite and aplite in metasediments and arnockites. Milestone 100 on Colombo–Deniyaya road. (Vertical section)

Offsetting does sometimes take place—but either in the wrong direction r simple opening or in a disproportionate amount, or both. Such fsetting is usually the result of early shearing movement, as for example Fig. 2D where offsetting has resulted from shearing and a quartz vein s been emplaced in the shear plane. This is in agreement with the servation made by King (1948, 463) who shows further that in some aces the rocks have undergone plastic deformation, the lateral displace- ents on either side of the dykes showing rapid variation both in magnitude d direction.

The absence of offsetting at the oblique intersection of two dykes as a iterion for replacement has recently been criticised (Wells & Bishop, 54, 104) on the ground that relative movement would negative the lational effect of injection. In the examples quoted above, and in fact in arly every instance examined by the author, there has always been a linear continuation of the earlier dyke across the later. If any horizontal movement has taken place in these examples it must have been of the exact amount in every one of them to compensate for the effect of dilation injection; as Goodspeed himself (1940) has remarked, this is too much of a coincidence to expect in every case. Wells & Bishop also point out that if the direction of movement is parallel to the earlier structure, the offsetting becomes zero. Here, too, not only is the coincidence too great, but there is no reason for relative movement to take place in this particular direction rather than in the innumerable other possible directions.

(b) Enclosure of Undisturbed Blocks of Invaded Rocks

The presence of undisturbed blocks of country-rock in granite is a common feature in migmatite areas, and such relics have been shown by several workers (Misch, 1949; Read, 1951) to be one of the strongest proofs of replacement or granitisation. Some pegmatites exhibit the same criteria (though on a much smaller scale) and in them the continuation of foliation can be followed undisturbed from wall-rock through relic and back into country-rock within a few feet, or even inches.

One of the best examples of this is seen in the Ittiwatana quarries, near milestone 11 on the Puttalam–Anuradhapura road. The country-rocks here are rapidly varying banded gneisses with narrow concordant pink granitic veins and crossed by pegmatites. Bands of different composition

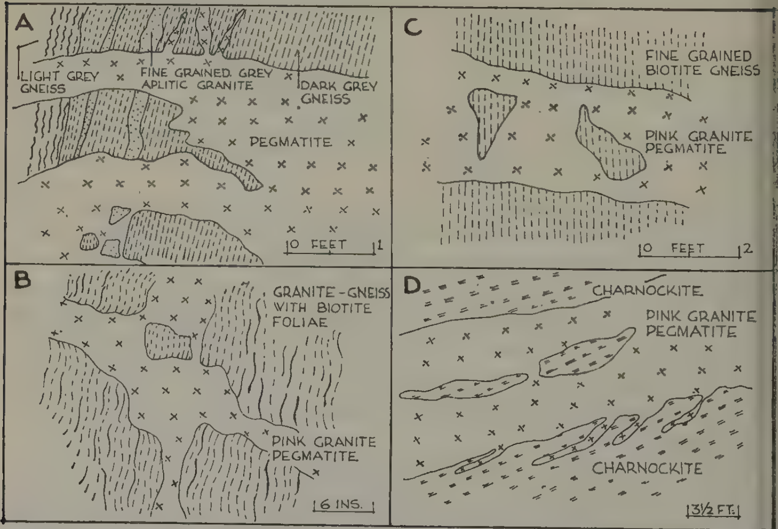


Fig. 5. Undisturbed relics of wall rock in pegmatite. A: Ittiwatana quarry, near milestone 11 on Puttalam–Anuradhapura road. B: Tabbowa. C: Tambutta, near Tonigala. D: Pattiwila quarry, Kelaniya, Colombo

the gneiss (Fig. 5A) can be followed into a large block of country-rock thin the pegmatite and, a few inches further, into small isolated patches country-rock in the pegmatite. Other examples are seen in Fig. 5B, C. In ttiwila quarry, Kelaniya, in the Colombo District (Fig. 5D), a large gmatite cuts obliquely across gneissic charnockite, and within it are und relic blocks of charnockite maintaining the same foliation as in the ll-rocks.

The writer's experience of gneissic relics in granitic rocks in the Pre-mbrian formations of Ceylon is that they are generally isolated bodies

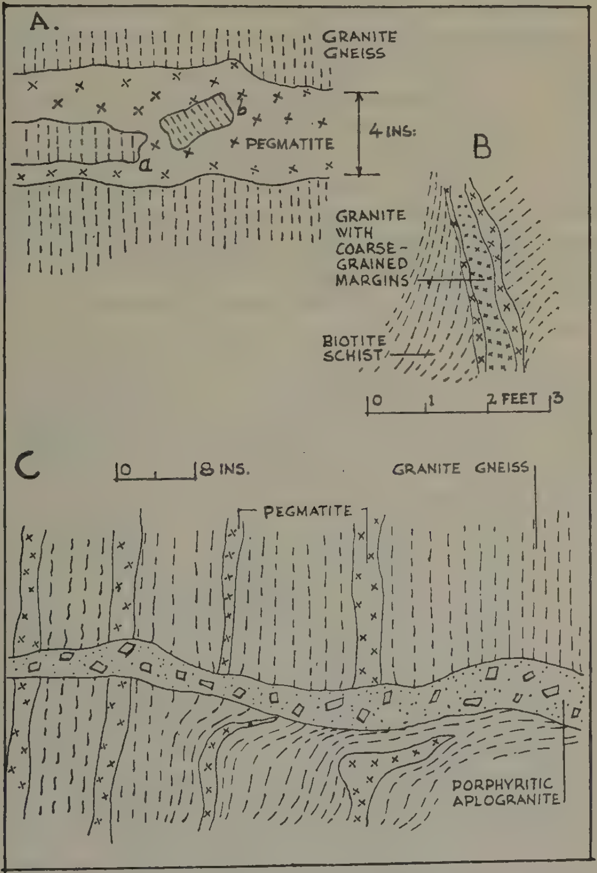


Fig. 6. A: Undisturbed relic block next to slightly disturbed block in cross-cutting pegmatite. Kaduwela Wewa, Hembarawa. B, C: Drag of foliation on one side of granite dyke. B: Nekettegama, Nikaweratiya. C: Bogahawewa, Wilgomuwa

of no great extent and that they are surrounded on all sides by the containing rocks. While it is still possible that the enclaves shown in Fig. 5A, B and C and Fig. 10A may have been attached in the third dimension, their irregular outlines makes such a possibility unlikely. It is hardly conceivable that large amounts of material could have been removed by the injection of such comparatively narrow dykes without there being the slightest trace of this material anywhere in the vicinity.

In some dykes movement does appear to have taken place, but such movement is generally local, and though certain parts of a dyke show no signs of disturbance, contiguous portions may present evidence of movement or plastic deformation. Thus, for example, in a dyke at Kaduwela Wewa, Hembarawa (Fig. 6A), one relic block is undisturbed but an inch or so away another block shows slight displacement. (The writer is aware that this figure may be interpreted in another way, viz. that one block was attached while the other was not, but considers this unlikely.) At Nekettegama, near Nikaweratiya (Fig. 6B), and at Bogahawewa, Wilgomuwa (Fig. 6C), are dykes in which only a portion of one wall shows movement, as indicated by the drag of the foliation of the gneiss at the contact. The felspar phenocrysts in the dyke in Fig. 6C do not show any flow structure and were probably formed after the localised movement. Goodspeed (1952, 358, pl. 2c) describes an almost identical phenomenon in a granitic

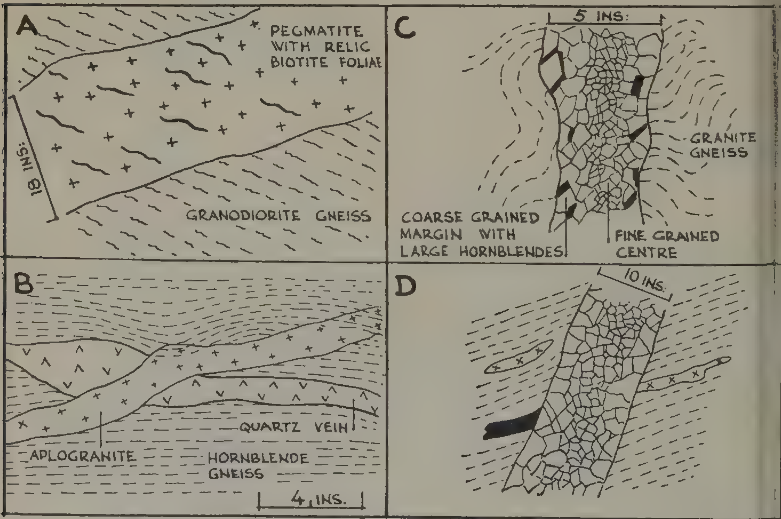


Fig. 7. A, B: Relic biotite foliae continuing undisturbed through cross-cutting dykes. A: Uriyawa Wewa, north of Tonigala. B: Panchalawewa, near Galgamuwa
C, D: Cross-cutting pegmatites with coarse-grained margins and fine-grained centres. C: Galgamuwa. D: Colombo

ke from Cornucopia, and has called such dykes 'mobilised replacement kes'.

(c) Continuation of Gneissic Foliae Through Dykes

n some localities thin biotite foliae or such elongate minerals as quartz g. 2A, inset), biotite flakes or hornblende crystals can be seen to extend oss the dyke or vein. Examples of this are shown in Fig. 7A, B, from the lgamuwa area, and the preservation of such delicate structures suggests t the dykes were formed metasomatically.

An early stage in such replacement can be seen in the Malabe and

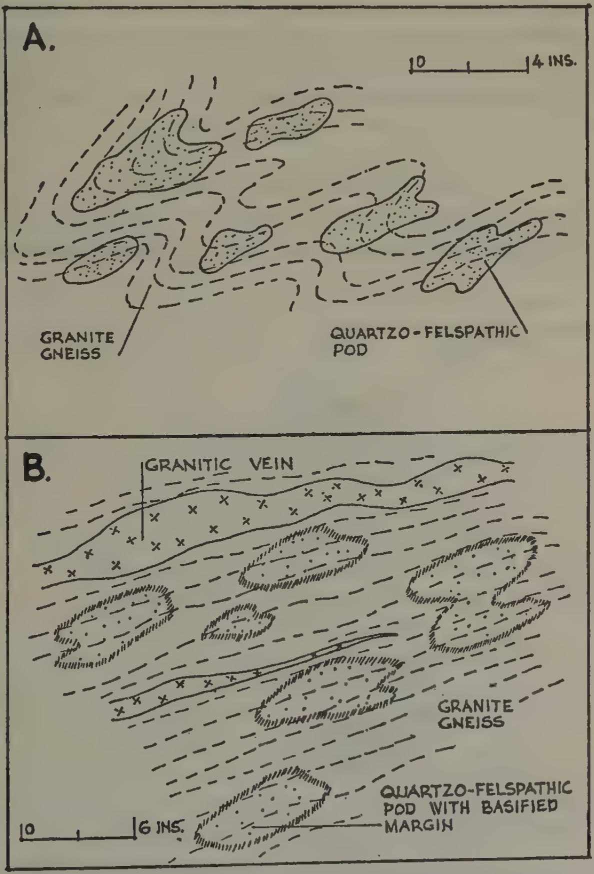


Fig. 8. A: Foliation of gneiss continuing through quartzo-felspathic pods. Malabe Quarry, Colombo. B: Mica-rich margins to quartzo-felspathic pods in gneiss. Malabe Quarry, Colombo

Arangala quarries near Colombo. Here, quartzo-felspathic pods with large felspars occur in biotite-gneisses whose foliation continues undisturbed through the pods (Fig. 8 A, B). Several such pods were seen to be surrounded by small basic selvedges composed of mica (Fig. 8 B). Elsewhere in the same quarries, several of these patches have coalesced to form pegmatite dykes or veins. Such evidence cannot of course be applied to every case but it does show one way in which pegmatites may be emplaced metasomatically. A similar phenomenon has been described in the granitic gneisses of W. Ardgour, Scotland (Harry, 1954, 294), where quartz-felspar bodies, frequently pegmatitic, form 'irregular patches, pods and thick bands'. These bodies are described as being margined by narrow zones formed by biotite enrichment, and are roughly parallel to the foliation of the containing gneiss; the only difference is that the biotites in the host rock do not pass through the pods but are 'deflected into parallelism with themselves'. Harry ascribes their formation to metamorphic segregation at favourable loci.

(d) Truncation of Basic Patches

Frequently, when dykes truncate gneisses with basic lenses (and pegmatite veins and lenses), the corresponding portions of the truncated rocks are

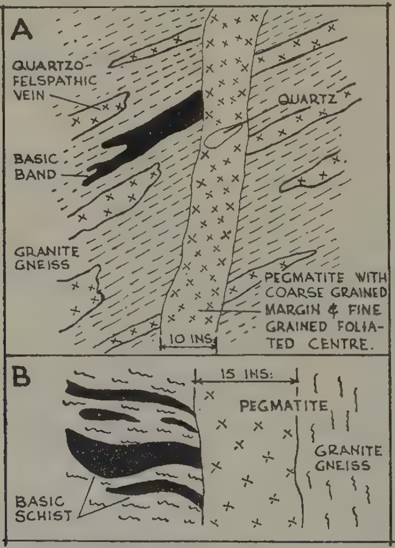


Fig. 9. A, B: Basic bands truncated by pegmatites; note lack of correspondence on either side of dyke.
A: Kaduwela quarry, near Colombo. (Vertical section). B: Kuda Tammanawa, Tonigala

sent either from the opposite side of the dyke or from within the dyke elf, where they would be expected if the dykes were emplaced by a lational mechanism.

A good example of this is seen at Kuda Tammanawa, near Tonigala, here a pinkish grey granite gneiss with basic schist lenses is truncated by nk granite pegmatites (Fig. 9B). The corresponding portions of the uncated basic lenses are completely missing. The same phenomenon is en at Kaduwela quarry near Colombo (Fig. 9A), where lack of offsetting the dyke intersections is present as well. It is specially noteworthy here at although the dyke is foliated parallel to the contacts, an elongated ns of vein quartz, such as is to be found in the surrounding gneiss, ntinues across the dyke in a direction parallel to the foliation of the all-rock. The foliation of the dyke has therefore not been produced by essure, but by some form of mimetic crystallisation along privileged ths thus leaving unaffected and undisturbed the quartz lens.

That relative movement can result in the concealment or removal of rresponding parts of truncated enclaves has been clearly shown by Wells Bishop (1954, 105). It is true to say, however, that in the hundreds of posures examined by the author, relative movement along dykes has rely been seen either in vertical or horizontal sections. The example own in Fig. 9A has in fact been taken from a vertical section and if any ovement has taken place it must have been a scissor-like movement nged at the bottom of the section.

(e) 'Pinch-and Swell' Structure of Dykes

One of the commonest features of granitic dykes in migmatised areas is eir extreme variation in width resulting in the so-called 'pinch-and-swell' ructures. In contrast, magmatic dykes emplaced dilationally, e.g. lerites, lamprophyres, etc., normally show a characteristically uniform idth. Such 'pinch-and-swell' structures, though commonest in concordant kes and veins, are sometimes found in those cutting obliquely across the liation.

An excellent example, seen on the road to Pandipanchawa in the algamuwa area, is shown in Fig. 10A, where well-foliated biotite gneiss cut obliquely by a pink pegmatite dyke. The pinches and swells have no stortional effect on the host rock and, in addition, an undisturbed relic the gneiss is found within the pegmatite; an early stage in the formation a similar relic block can be seen at the top of the figure. The presence of combination of these features demands a replacement process and not lation-injection as the mode of formation of the dyke. In a dyke from alboda Estate, Rangala (Fig. 10B), the 'pinch' in the dyke coincides with quartzose band in the gneiss, suggesting that the acid band is more sistant to replacement than the rest of the rock (Read, 1951).

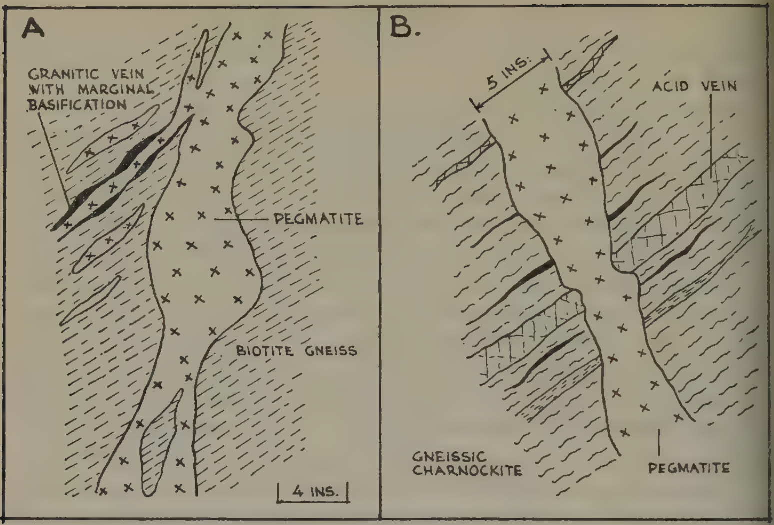


Fig. 10. A: Pinch-and-swell structure in pegmatite containing undisturbed relic of wall rock; note incipient relic formation at top of dyke. Pandipanchawa, Galgamuwa. B: Coincidence of pinch in pegmatite dyke with quartzose band in gneiss. Galboda Estate, Rangala

(f) Coarse-grained Margins and Finer-grained Cores

One of the diagnostic and characteristic textures of igneous dykes emplaced by dilation-injection is the presence of chilled or fine-grained edges and coarser-grained centres. Analogous textures are very rarely observed in the pegmatite and aplites of the areas under review. On the contrary, it is common to see such dykes and veins with coarse-grained margins and finer-grained cores, as for example in dykes from the Galgamuwa and Colombo areas (Fig. 7C, D).

(g) Sharp and Diffuse Contacts: Reaction Zones

Non-dilational dykes and veins can show either sharp contacts or gradational contacts with the country-rocks (Goodspeed, 1940, 194) and reaction zones (generally of basic enrichment) are commonly present. Dilation-injection dykes on the other hand have characteristically sharp contacts with the country-rocks, and reaction zones are rare.

In the areas under review, the dykes and veins show both types of contacts. Where they are gradational it is often extremely difficult to demarcate the margin between dyke and country-rock, and this is especially so in the pegmatitic dykes of the Tonigala granite. Even apparently sharp

ntacts can be deceptive and may be due to a colour difference (Good-
ed & Coombs, 1932), or to the presence of a reaction zone in the wall-
k (King, 1948, 474 and fig. 19). The 'purple' dykes in the Tonigala area
oray, 1958) form good examples. In thin section, such 'sharp' boun-
ries are seen to be in fact gradational.

3. MICROSCOPIC FEATURES

The most marked textural feature of most of these dykes and veins is the
sence of a clear demarcation between the minerals of the dyke and those
the wall rock. In nearly every dyke-margin examined there was a
ring of individual crystals at all points along the contacts. For example,
a thin section (K. 540)[1] cut across the margin of a pink granite dyke in
tite gneiss from Galgamuwa, one can see minute biotite flakes which
ntinue to the edge of the dyke and even protrude into it without being
torted. The actual contact is a very irregular line. The same can be
served in thin section (K. 502) from the Kaduwela quarry near Colombo.

The dykes and veins discussed in this paper, unlike the dilation dykes
ich generally have glassy borders and a progressive increase of grain
e towards the centre (Goodspeed, 1940), have a consistently irregular
ture throughout. The texture is typically crystalloblastic with the
ividual crystals interfering with each other; crenulate margins are
nspicuous. In nearly every section, lobes of myrmekite on the fringes of
gioclase crystals can also be seen encroaching on the potash felspars.

An interesting example of a reaction zone between gneiss and pegmatite
s observed in the slide (K. 963A) from Ittiwatana quarry (p. 78). The
eiss is composed of quartz, plagioclase, hornblende (sodic on the mar-
s), diopside and sphene and the margin of the dyke is at right angles to
foliation. The amphibole in the gneiss decreases and finally disappears
the contact is approached; but there is a simultaneous increase both in
amount and size of pyroxene, sphene and ores, some crystals of
roxene being about 2 mm. in length. The pegmatite is a coarse-grained
crocline–plagioclase–quartz rock in which the potash felspar is being
laced on the edges by numerous lobes of myrmekitic plagioclase. A few
ic grains of green pyroxene exactly similar to those in the reaction zone
present in the pegmatite. The contact is extremely irregular, the
roxene of the wall-rock protruding conspicuously into the dyke without
ng disturbed. Such reaction zones are seldom, if ever, seen in magmatic
ection dykes, but they are common in those formed by replacement.
re the reaction zone appears to be due to the metasomatic addition of
, Mg and Fe to the wall-rock of a calc gneiss.

Figures refer to the Rock and Thin Section Collection of the Department of Mineralogy
ological Survey of Ceylon), Colombo.

4. MODE OF EMPLACEMENT

Some of the dykes examined by Goodspeed (1940) showed several transitional stages from alteration along a single fissure to a complete replacement dyke. This was strikingly so in dykes occurring in hornfelses. Selective replacement along several closely spaced fractures with relic septa and foliae of the country-rock was followed by the merging of parallel bands with long thin relic inclusions parallel to the walls of the dyke. The disappearance of all original material was the final stage.

In the dykes and veins of the Pre-Cambrian areas reviewed in the foregoing pages, such features have only rarely been noted, as for example in the 'purple' dykes of the Tonigala area (Cooray, 1958), either because they have been overlooked, or because replacement and recrystallisation have proceeded to such an extent that such transitional stages have been completely obliterated. It should be mentioned, however, that relic folia of country-rock parallel to the walls are common enough in *concordant* pegmatite dykes and veins in these migmatised areas.

The evidence presented above appears, however, to be conclusive that in the examples discussed, which incidentally are typical of most Pre-Cambrian areas of Ceylon, dykes and veins have been emplaced non-dilationally, and therefore by metasomatic replacement. Further, as Misch (1949) has shown, the preservation of undisturbed relic structures and minerals in the dykes traversing the Ts'angshan gneisses is a feature of post-kinematic static replacement. It is therefore considered that, in Ceylon, the emplacement of cross-cutting dykes and veins of pegmatite and aplite in the migmatites and granites of the Vijayan Series took place after the main episodes of granitisation, by replacement and along fractures in the consolidated rocks. Although the evidence on which this deduction is based is taken from a large area, it is all consistent with such a mode of emplacement and no contradictory evidence has so far been met with.

5. PTYGMATIC VEINS

The term 'ptygmatic' originally coined by Sederholm is now used to describe the tortuous quartzo-felspathic veins which occur in areas of granitisation. Several modes of formation have been suggested in the past and Wilson (1952) has summarised these views as follows:

(i) *secondary* folds which result from a passive response to movements of the host rock after the emplacement of the veins;

(ii) *primary* folds in which the tortuosities are the result of injection under special conditions.

Examples of both types have been met with in the Pre-Cambrian areas of Ceylon examined by the author. An example of the first type, viz.

ondary fold, is seen on the recently constructed Galgamuwa–Palukada-
la road where a ptygmatic vein of pink granite pegmatite traverses
meated biotite gneiss (Fig. 11 A). The vein shows all the features which
uld be expected from the compression of a competent planar body in an
ompetent body of rock, resulting in a combination of plastic deforma-
n and intense folding similar to that demonstrated experimentally by

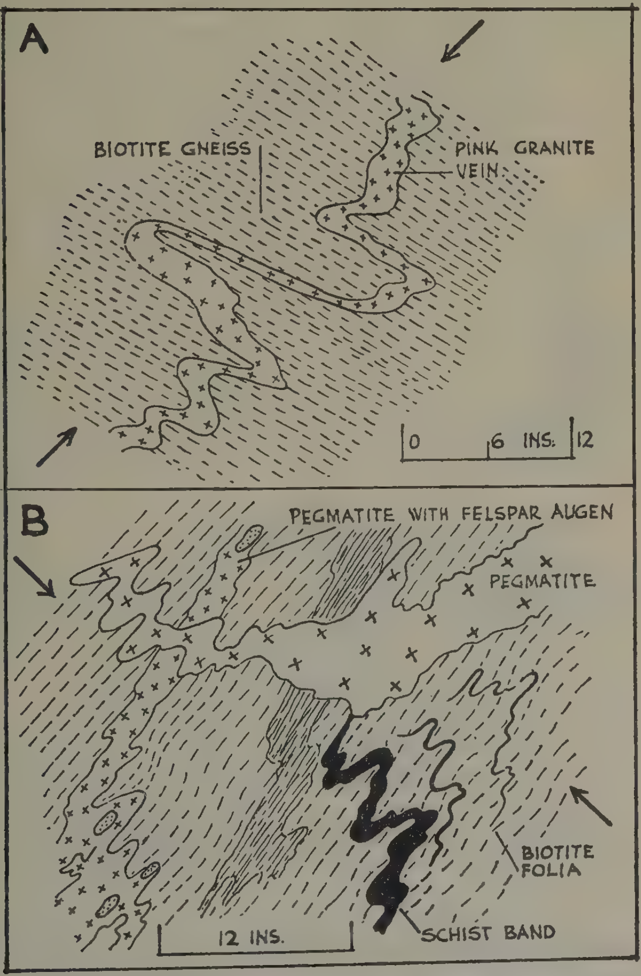


Fig. 11. A: Ptygmatic granite vein in biotite gneiss. Galgamuwa–Palukadawela road. B: Pegmatite veins, basic schist bands and biotite foliae showing ptygmatic forms. Ittiwatana quarry, near Puttalam

Keunen (1938). These features are the parallelism of the foliation of the host rock to the axial planes of the plications and the thinness of the vein in a direction parallel to the maximum distension and its comparative thickness in the direction of maximum compression.

Similar features are seen in Fig. 11 B from the Ittiwatana quarries near

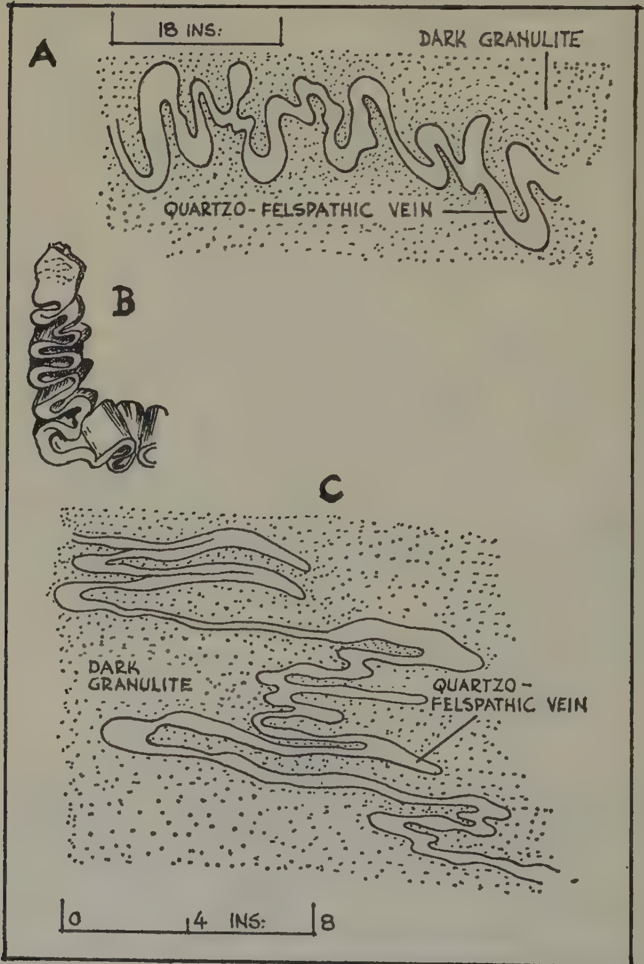


Fig. 12. A, C: Ptygmatic vein in unfoliated granulite. Mayilewa, on Galgamuwa–Maha Nanneriya road. B: Structure developed experimentally by injecting ribbon of putty into soft jelly. (After Wilson, 1952, fig. 6 c)

ttalam where pegmatites as well as biotite schist lenses and foliae are ded into ptygmatic forms.

Examples of the second type, viz. primary folds, are conspicuously splayed in the Galgamuwa area where they are confined to a band of rk grey quartz–plagioclase–microperthite–hornblende–biotite granulite. ey are best seen in two quarries near Mayilewa on the Galgamuwa–aha Nanneriya road (Fig. 12A, C). The granulite is unfoliated and nost massive except for the presence of a few narrow parallel leucocratic ins in it. The form of the veins is strikingly like the structures which ilson (1952, fig. 6) found were developed experimentally by injecting a bon of putty into soft jelly. Wilson's fig. 6c is reproduced here for mparison (Fig. 12B). According to him these primary 'ptygmatic ections' are formed when the country-rock at the time and place of the ection is more mobile than the magmatic material of the veins themselves. e contortions are formed by the plastic buckling of the vein due to gmatic pressure from behind, with concomitant displacement of lding contiguous host rock.

It is reasonable to believe that the examples shown here (Fig. 12A, C) re formed in a similar manner, the difference in the form of the two veins ing due either to the greater mobility of the host rock, or to a partial llapse of the vein, or to a combination of both. It is perhaps significant at the 'ptygmatic injections' are here confined to a rock with a direction-s fabric in an area of foliated gneisses and granitic gneisses.

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The Cenomanian Deposits of South-East Devonshire

The Cenomanian Limestone and Contiguous Deposits West of Beer

by WILLIAM E. SMITH

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BSTRACT: The distribution, lithology and regional variations in the thickness of the enomanian deposits in south-east Devon are summarised. Three distinctive facies are cognised, viz. (i) the Lower Chalk facies, (ii) the Cenomanian Limestone facies, and i) the calcareous sand or Wilmington facies.

A detailed description is given of the lateral variations in the lithology, thickness and certain palaeontological features of the cliff-sections and inland exposures of Cenomanian rocks west of Beer. Between Hooken Cliff and Branscombe West Cliff the lateral variations in the lithology and thickness of the Cenomanian rocks are related to the Branscombe Mouth Ridge. This was a submarine ridge which periodically developed in the Branscombe Mouth region as a result of anticlinal earth movements along an elongated pericline, possibly of asymmetric type, and with an axis elongated in a general N.–S. direction. The greatest uplift along this axis occurred in late Cenomanian times and led to the formation of a ridge that was not buried beneath Middle Chalk deposits until after the deposition of the *Inoceramus labiatus* Zone. The anomalous fauna of the *Inoceramus labiatus* Zone west of Branscombe Mouth is ascribed to the influence of the Branscombe Mouth Ridge.

The thick Cenomanian deposits at the eastern end of Hooken Cliff accumulated in a tectonically stable area or depression lying between the Branscombe Mouth Ridge and the Beer Beach–Little Beach Ridge (Smith, 1957a). They display lithological features indicative of current activity that was generally weaker than that responsible for vigorous erosion over the adjacent ridges. The thick Cenomanian deposits in the Bovey Lane sand-pit, about one mile north-west of Beer, include beds of Wilmington facies and appear to have accumulated in another depression connected with the northerly plunge of the Branscombe Mouth Ridge. The lateral variations in the thickness of the Cenomanian Limestone west of Branscombe Mouth are mainly due to the presence of undulating erosion-surfaces, including comparatively deep scour-channels in the top of the Upper Greensand. The thin and incomplete development of the Cenomanian Limestone at Donkey Linhay may, however, be indicative of small-scale intra-Cenomanian anticlinal earth-movements in that region.

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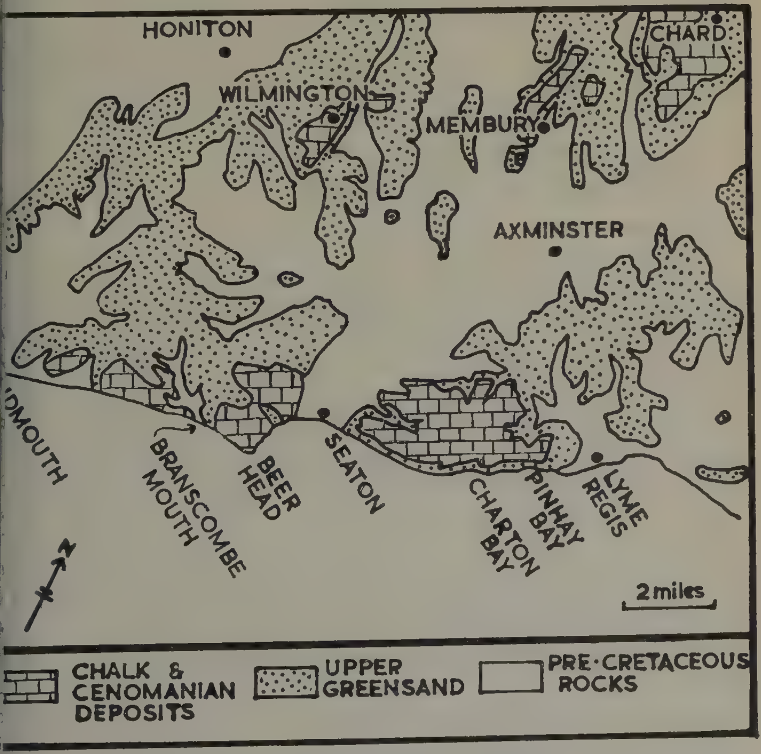
1. INTRODUCTION

(a) The Distribution of the Cenomanian Deposits

THE CRETACEOUS rocks of south-east Devon form the southern part of a highly dissected upland region or plateau which extends northwards from the coast between Lyme Regis and Sidmouth and culminates in the Blackdown Hills of Somerset. This plateau rises to a little over 1000 feet in the Blackdown Hills but descends seawards to between 400 and 500 feet along the coast. Most of the high ground in the region is capped by Upper Greensand and drift deposits but outliers of Chalk and Cenomanian deposits are locally preserved, especially on the downthrow side of important faults (Fig. 1). The largest of these outliers extends for about two miles north of the coast between Haven Cliff, east of Seaton, and Ware Cliff, west of Lyme Regis. Another important outlier extends for a similar distance northwards from the coast between Whitecliff, west of Seaton, and Branscombe East Cliff. This outlier occurs on the western or downthrow side of an important fault which throws the Cretaceous deposits against the Keuper Marl; it includes some of the youngest Chalk in the region preserved as a result of synclinal folding in the Beer district. Cenomanian and Turonian deposits also cap the high ground terminating in the cliffs west of Branscombe Mouth, the westernmost development of these deposits being that on Maynard's Cliff (30/150878), about one and a half mile east of Sidmouth.

he Cenomanian deposits of the Wilmington district, about seven miles
th of Branscombe Mouth, have long been famous for their rich fauna.
se deposits, together with part of the Middle Chalk, are preserved as an
lier on the western or downthrow side of an important N.–S. fault which
ngs up the Keuper Marl at the eastern end of Wilmington village.
nomanian rocks also occur on the upthrow side of this fault and are
osed in a completely decalcified condition in a sand-pit alongside the
e leading from Home Bush to Watchcombe (30/218977). It is also
sible that Cenomanian deposits occur beneath the drift cover on
ckland Hill, north-east of Wilmington. The lateral extent of these more
terly Cenomanian outliers is at present uncertain but it is clear that they
small in size and restricted to the high ground immediately east of the
mington Fault.



1. Geological sketch-map showing the relative positions of the coastal Chalk
tliers between Lyme Regis and Sidmouth and those in the Wilmington and Membury
tricts

Cenomanian deposits are present below the Middle Chalk in an outlier north-west of Axminster, close to the Devon–Somerset border. This outlier is elongated in a N.–S. direction and passes through Membury, a village situated about four and a half miles north-east of Wilmington and about eight and a half miles north of Haven Cliff, Seaton. The Cretaceous rocks are preserved on the western or downthrow side of an important N.–S. fault which passes just east of Membury village and brings up the Rhaetic and Lower Lias on its eastern side. The outlier is one of a group lying for the most part outside the borders of Devon, in the Chard and Crewkerne districts of Somerset; those outside the borders of Devon are not dealt with in the present account.

(b) Lateral Variations in the Lithology and Thickness of the Cenomanian Deposits

A remarkable feature of the various outliers of Upper Cretaceous rocks in south-east Devon is that, despite their close proximity, they present three distinct developments of the Cenomanian Stage, viz.

(i) the Lower Chalk facies,
(ii) the Cenomanian Limestone facies,
(iii) the calcareous sand or Wilmington facies.

(*i*) *The Lower Chalk Facies.* In the Membury outlier the Cenomanian is represented by at least fifty feet of white or cream-coloured chalk containing a few layers of white, porcellaneous, nodular flint. The lowermost three or four feet of this chalk is sandy and glauconitic, the amount of sand and glauconite increasing downwards. The base of the sandy, glauconitic chalk rests upon a thin, conglomeratic bed containing numerous fossils, many of which are phosphatised, and abundant brown phosphatised pebbles of limestone sometimes encrusted with small unphosphatised fossils such as small serpulids. This bed is regarded as the Chalk Basement-bed since it contains Cenomanian ammonites and rests directly upon an erosion-surface cut across the hard, calcareous sandstone which is generally taken to be the top of the Upper Greensand. This development of the Cenomanian in the Membury outlier is clearly an extension of the Lower Chalk of the Chard and Crewkerne districts of Somerset and of that in southern Dorset.

(*ii*) *The Cenomanian Limestone Facies.* In the coastal outliers between Lyme Regis and Sidmouth the Cenomanian is not present as a thick deposit of chalk but is represented by a few feet of indurated, sandy limestones collectively known as the Cenomanian Limestone. There are no

all outliers serving to link the Cenomanian deposits at Membury with
se on or near the coast, post-Cretaceous erosion having completely
noved deposits which could have provided direct evidence of the
ationship between the Lower Chalk and the Cenomanian Limestone. It
lear, however, that the lateral change in the development of the Ceno-
nian must have been very rapid since the Lower Chalk at Membury is
ly about eight miles from the nearest occurrence of the Cenomanian
nestone.

The Cenomanian Limestone is not only much thinner than the Lower
alk at Membury, but itself undergoes rapid lateral changes in thickness.
at least one place, viz. at the extreme western end of Charton Cliff, the
mation is absent and the Middle Chalk rests directly upon the Upper
eensand. In most of the coastal sections the thickness ranges from one
four feet, but in a few places, e.g. at Hooken Cliff, west of Beer Head, and
Beer Head itself, it is much greater, the thickness at Hooken Cliff being
arly thirty feet. The Cenomanian Limestone consists of at least four
tinct beds, the relative thicknesses of which vary considerably from place
place. The variation in the overall thickness is often due to the wedging-
t of one or more of the beds or to sudden lateral increases in the thick-
ss of one or more of them. Sections showing the superposition of all four
very rare and occur only at localities, such as Hooken Cliff, where the
mation as a whole is exceptionally thick. In all sections the bed forming
local base of the formation rests abruptly upon a well-defined erosion-
ne at the top of the Upper Greensand. Erosion-planes, frequently
crusted with a brownish phosphatic veneer, also occur within the forma-
n and, together with thin layers of phosphatised pebbles of limestone,
ve as boundaries between the various divisions.

The four beds comprising the Cenomanian Limestone are conveniently
med according to the scheme proposed by Jukes-Browne in his descrip-
n of the cliff-sections in south Devon (1903, 130–44). The principal
tinguishing features used by the present writer in identifying the various
ds are given below:

SUBDIVISIONS OF THE CENOMANIAN LIMESTONE IN SOUTH-EAST DEVON

PERMOST NOMANIAN LOWERMOST RONIAN	**Division C.** A white, chalky deposit containing abundant sand and glauconite and scattered dark brown phosphatic pellets, passing up, without a break, into typical Middle Chalk but resting abruptly upon Division B. Fossils abundant, one of the most characteristic being *Orbirhynchia wiesti* (Quenstedt). A bed of very local distribution, best developed in fallen blocks at the eastern end of the Hooken Landslip, in the cliff at the rear of this landslip and in fallen blocks on the beach at Humble Point, at the eastern end of Charton Bay. (Thickness 0–7 feet.)

—FAUNAL BREAK—

LOWER CENOMANIAN

Division B. A hard, white, sandy limestone usually containing conspicuous grains of glauconite. Locally (e.g. at Hooken Cliff) the sand and glauconite are concentrated into elongated patches between which the limestone contains very little detritus and resembles indurated chalk. Green or brown streaks and green- or brown-stained pebbles of limestone are common, especially towards the top. The top is usually an erosion-surface but is occasionally a thin layer of brownish, phosphatised pebbles of limestone. Fossils are not common, except among the phosphatised pebbles of limestone where ammonites, indicative of more than one horizon in the Cenomanian, may be abundant, as at Humble Point. *Holaster subglobosus* Leske appears to be restricted to this division. Division B is one of the most persistent beds in the Cenomanian Limestone. (Thickness 0–5 feet.)

Division A. A hard, white, sandy limestone containing numerous shells and shell-fragments which are often silicified. Glauconite is very scarce and never conspicuous, pebbles of limestone are usually rare, except towards the top. The top is usually a well-defined, brown-stained erosion-surface welded on to the base of Division B, the base is equally well defined and rests abruptly upon an erosion-surface at the top of the Upper Greensand. Crevices filled with early Cenomanian deposits are locally present at the top of the Upper Greensand.

Two subdivisions can be recognised, the lower, Division A 1, contains abundant grains and small pebbles of quartz and abundant large coral-like polyzoa, *Ceriopora ramulosa* Michelin. This subdivision is mainly restricted to the thicker developments of the Cenomanian Limestone such as occur at Beer Head and Hooken Cliff. The large polyzoa do not normally occur in the upper subdivision, Division A 2, which is a shelly limestone containing comparatively little sand. Division A 2 is much more widespread than the lower subdivision and often rests directly upon the Upper Greensand. The boundary between the two subdivisions may be transitional and ill-defined or marked by a sharp lithological change or an erosion-surface. Fossils are common in both subdivisions but are often difficult to extract. (Total thickness 0–18 feet.)

(*iii*) *The Calcareous Sand or Wilmington Facies.* The third distinctive lithological facies of the Cenomanian in south-east Devon is a coarse calcareous sand consisting predominantly of grains and small pebbles of quartz enclosed within a white, powdery, chalky matrix. In regions where the cover of Middle Chalk has been long removed the sands are usually decalcified and the chalky matrix completely absent. When the cover of Middle Chalk is still present the chalky matrix of the uppermost few feet of sands is usually more coherent than that of the underlying deposit, a variable thickness of hard calcareous sandstone known as ‘grizzle’ being often present at or near the top. The Cenomanian sands or sandstones are usually separated from the Middle Chalk by a small thickness of glauconsandy limestone resembling Division B of the Cenomanian Limestone e coastal sections.

his sandy facies of the Cenomanian is undoubtedly more widespread previous records suggest. It will be shown later in this account that paratively thick developments of Cenomanian sands are not only erved as outliers on the high ground east of the Wilmington Fault but similar sands are also present over a restricted area in the Beer district well exposed in a small sand-pit alongside Bovey Lane, Beer (30/217900). e their development is particularly important since they occur within utlier in which the Cenomanian Limestone is the more typical facies.

ossils, including typical Lower Cenomanian ammonites, are usually ndant in the few feet of more or less coherent strata immediately w the Middle Chalk but are very rare in the main mass of the calcareous ds and in the completely decalcified deposits. This scarcity of fossils in major part of the sands and the almost complete absence of exposures eir basal part hinder the precise determination of their stratigraphical ge. In a badly overgrown sand-pit on the north side of Wilmington tchin's sand-pit 31/212003) a shell-bed, crowded with casts of *Trigonia* containing occasional Lower Cenomanian ammonites, underlies the areous sands and rests abruptly upon a hard calcareous sandstone mbling the top of the Upper Greensand (Smith, 1957b, 152). At this lity the entire thickness of the calcareous sands can therefore be rred to the Lower Cenomanian with confidence. The lateral extent of shell-bed and the precise relationship between the calcareous sands and Upper Greensand in the area south of Wilmington are at present nown. As a result considerable thicknesses of calcareous sands are rred to the Cenomanian because they pass up without a break into siliferous sandstones of undoubted Lower Cenomanian age; clearly a nomanian age for the whole of the calcareous sand facies cannot be arded as firmly established.

he scarcity of exposures in the basal sands prevents an accurate ermination of any lateral variations in thickness such as occur in the nomanian Limestone. Jukes-Browne (1903) estimated that the arena- us Cenomanian strata are about forty feet thick in the White Hart d-pit (30/208999), at the western end of Wilmington village. The sands uld appear to thin towards the north-east, however, since in Hutchin's d-pit, about one-third of a mile north-east of the White Hart pit, the e of the Middle Chalk is less than twenty feet above the presumed ement-bed. This north-easterly attenuation of the sands is particularly nificant since they are completely absent from the Membury outlier, out four miles farther north-east. Furthermore, the thinning-out of the ds is accompanied by the development of a considerable thickness of wer Chalk, a formation which is not present as a lithological facies in the Wilmington district. The great contrast between the developments of the Cenomanian at Wilmington and Membury is one of the more remarkable features exhibited by the Cretaceous rocks investigated by the author.

(c) Summary of Previous Work

The earliest description of the Cretaceous rocks of south-east Devon is that given by De la Beche (1826, 1839). He gave a very brief description of the Chalk and 'greensands beneath the Chalk' in the Beer district and between Pinhay and Dowlands, west of Lyme Regis. The Cenomanian Limestone is clearly approximately equivalent to the formation termed 'Chalk with quartz grains' by De la Beche. It would appear, however, that he included beds now grouped as the lower part of the Middle Chalk since the thicknesses of 'Chalk with quartz grains' given by him for sections near Lyme Regis are considerably greater than the maximum thickness of the Cenomanian Limestone in these sections.

The first detailed description of a small part of the coastal section is that given by Meyer (1874) who divided the sequence of Cretaceous rocks exposed in the Beer district into twenty numbered subdivisions, for each of which he gave a faunal list; his subdivisions 10 to 13 cover the strata classed as Cenomanian Limestone in the present paper. The earliest application of the zonal concept to the subdivision of the Cretaceous rocks in the cliffs of south-east Devon was made by Barrois (1876). He recognised the Cenomanian age of the beds comprising the Cenomanian Limestone but also included in this Stage a considerable thickness of beds forming the upper part of the Upper Greensand and normally classed as Upper Albian by British geologists (Jukes-Browne, 1896; Spath, 1943, 743–5). Barrois noted the erosion-surface at the top of the Cenomanian Limestone and the palaeontological break at the base of the Middle Chalk, and suggested that both were due to earth-movements during Cenomanian times.

The most important contributions to our knowledge of the Cretaceous rocks of south-east Devon were made by Jukes-Browne (1896, 1900, 1903, 1904). He not only gave detailed descriptions of all the coastal exposures between Lyme Regis and Sidmouth but also gave the earliest systematic descriptions of most of the inland sections, including those in the Wilmington and Membury districts. Jukes-Browne did not use the formation-name 'Cenomanian Limestone' but used the term 'Arenaceous Beds' in which he recognised four divisions in the coastal sections, viz. Divisions A1, A2, B and C. Divisions A1 to B were grouped with the calcareous sands at Wilmington as the Zone of *Ammonites mantelli* and classed as Lower Cenomanian. Division C was shown to be considerably younger and separated from the underlying beds by a pronounced faunal break. He suggested that it was of very late Cenomanian age and probably equivalent to the Subzone of *Actinocamax plenus* but might be a passage-bed linking

e Cenomanian and Turonian. In his monographic description of the
nglish Chalk, Jukes-Browne (1903, 1904) published lengthy faunal lists
om the various Cenomanian deposits in south-east Devon and in the
rlier volume gave petrographic descriptions of a small number of
mples, including lists of some of the detrital minerals contained in them.
e also attempted to explain the lateral variations in the thickness and
hology of the Cenomanian deposits in terms of variations in the distribu-
on and intensity of marine currents capable of inhibiting deposition and
eroding previously deposited sediments. In a brief footnote (1903, 351)
further suggested that differential (unequal) penecontemporaneous
rth-movements may also have been a factor.

Very little additional information has been published since 1904. The
scriptions of the Cretaceous rocks given in the *Geological Survey*
emoir relating to the country near Sidmouth and Lyme Regis (Wood-
ard & Ussher, 1911) and *British Regional Geology*—'*South-West*
ngland' (Dewey, 1958) are based upon the observations made by Jukes-
rowne. The present author has already published a detailed description of
e lateral variations in the Cenomanian Limestone exposed in the cliff-
ctions between Whitecliff and the Hooken Landslip, west of Beer Head.
he clarity, continuity and easy accessibility of the exposures of the
enomanian Limestone in these sections, combined with the absence of
tensive landslips and the presence of major right-angled bends in the
ast-line, have permitted a three-dimensional interpretation of the
ructures in the Cretaceous rocks. It has been shown that small-scale
termittent intra-Cenomanian earth-movements were responsible for *local*
riation in the thickness of the Cenomanian deposits which were preserved
ter regional current-scour. In the Beer district the Cenomanian Lime-
one is thinnest and least complete over the crest of a small, elongated,
tra-Cretaceous pericline passing from the west side of Beer Beach to
ittle Beach, immediately west of Beer Head. This pericline was the site of
termittent small-scale tectonic activity which occurred before and during
enomanian times and lasted at least into early Turonian times.

The present paper completes the description and interpretation of the
teral variations in the thickness, lithology and certain faunal character-
tics of the Cenomanian deposits in the outliers west of Seaton. Subsequent
rts will deal with Cenomanian deposits elsewhere in south-east Devon,
d with the sedimentary petrography and sedimentation-history of the
enomanian deposits of south-east Devon as a whole.

2. SECTIONS AT THE EASTERN END OF THE HOOKEN LANDSLIP

The Cenomanian Limestone is comparatively thick at Beer Head but
ins as it is traced through the cliff at the rear of Little Beach. Fallen

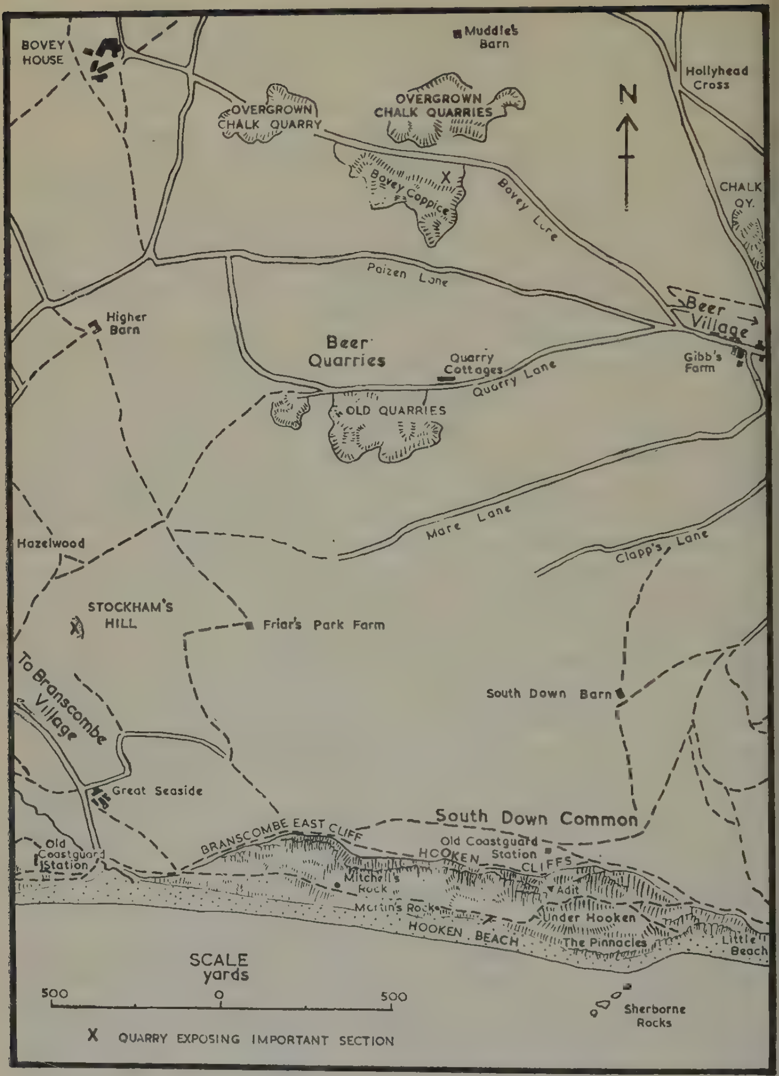
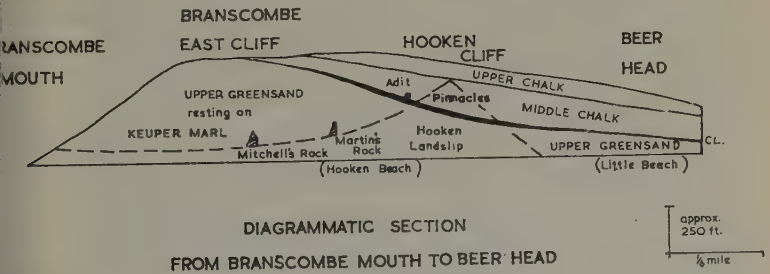


Fig. 2. Locality-map of the area between Beer and Branscombe, SE. Devon

blocks at the western end of the beach show, however, that it thickens again westwards (Smith, 1957a), and this increase in thickness continues into the eastern end of Hooken Cliff. Both here, and in fallen blocks in the adjacent landslip, the Cenomanian deposits are thicker than in any other ection along the coast of south-east Devon. The increase in total thickness s partly due to the incoming of Division C but is also due to increases in he thickness of both Division A and Division B. These thicker developnents of the two lower divisions show some important lithological ifferences from their equivalents in the sections between Whitecliff and Little Beach. There are three sections exposing the whole or a large part of he Cenomanian sequence:

a) A Large Foundered Block Forming the most Southerly Projection at the Western End of Little Beach (30/221879)

This block provides the clearest and most accessible section since it lies within reach of storm-waves and has come to rest with a steep landward tilt so that each bed can be closely examined without climbing. The basal part of the Cenomanian Limestone is, however, at present buried beneath beach-shingle; the total thickness cannot therefore be determined but must exceed twenty feet.



DIAGRAMMATIC SECTION

FROM BRANSCOMBE MOUTH TO BEER HEAD

Fig. 3

Division C. Soft chalky limestone containing scattered brownish phosphatic pellets and abundant sand and glauconite, not concentrated into lenticular streaks. Passes upwards into normal Middle Chalk by progressive decrease in the frequency of detrital grains; there is no well-defined upper surface and any thickness ascribed to the deposit is somewhat arbitrary. The thickness quoted is based on the fact that chalk with a lithology and faunal content comparable with that of the *Inoceramus labiatus* Zone elsewhere is first encountered at about four feet above the top of Division B and is further characterised by the common occurrence of *Orbirhynchia*, especially *O. wiesti* (Quenstedt). (About 4 feet.)

Division B (Fig. 4a). A thin bed containing numerous green- and brown-stained phosphatised pebbles of limestone overlain abruptly by Division C. (1 to 2 inches.)

Rough, nodular, sandy glauconitic limestone. (About 1 foot.)

Hard white limestone containing lenticular streaks rich in grains of quartz and glauconite several inches in length and half an inch or more thick. The streaks are almost always elongated laterally parallel to the top of the bed

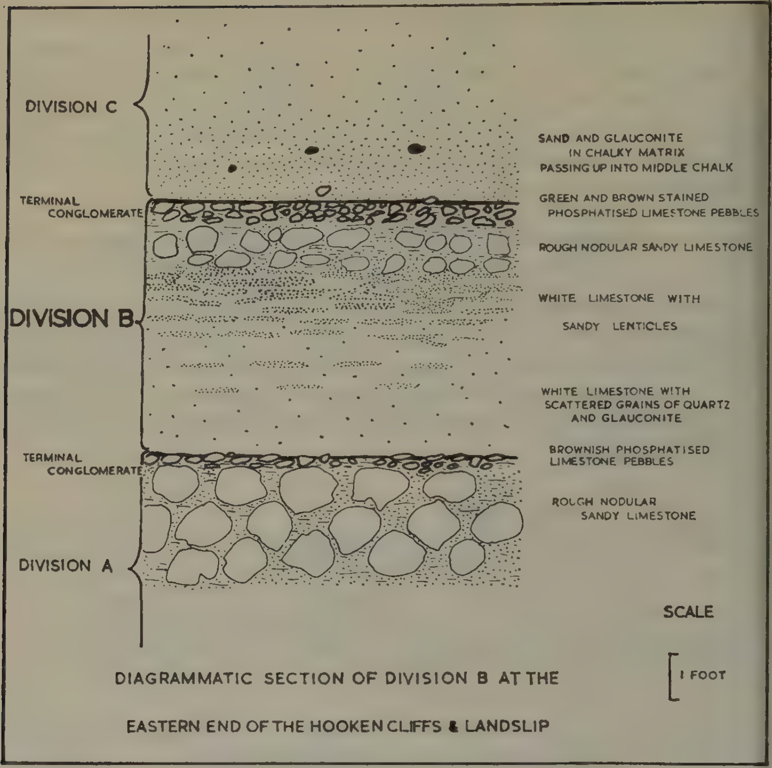

DIVISION C
TERMINAL CONGLOMERATE
DIVISION B
TERMINAL CONGLOMERATE
DIVISION A
SAND AND GLAUCONITE IN CHALKY MATRIX PASSING UP INTO MIDDLE CHALK
GREEN AND BROWN STAINED PHOSPHATISED LIMESTONE PEBBLES
ROUGH NODULAR SANDY LIMESTONE
WHITE LIMESTONE WITH SANDY LENTICLES
WHITE LIMESTONE WITH SCATTERED GRAINS OF QUARTZ AND GLAUCONITE
BROWNISH PHOSPHATISED LIMESTONE PEBBLES
ROUGH NODULAR SANDY LIMESTONE
SCALE
1 FOOT
DIAGRAMMATIC SECTION OF DIVISION B AT THE EASTERN END OF THE HOOKEN CLIFFS & LANDSLIP

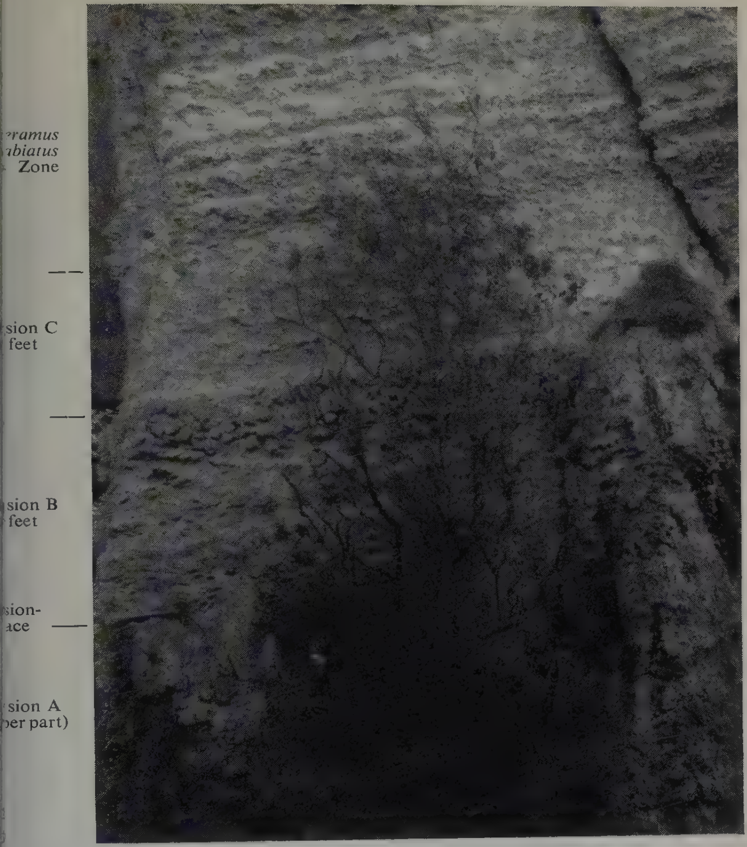
Fig. 4a

although occasionally splitting into two or more transgressive branches. At lower levels they are widely spaced but the vertical distance between adjacent streaks gradually decreases upwards and towards the top they appear to locally coalesce with underlying or overlying streaks so that masses of limestone sometimes appear to be completely enveloped by glauconitic sandstone. (About 2½ feet.)

Hard white limestone resembling indurated chalk and containing very little sand or glauconite, more or less uniformly distributed in the calcareous matrix. A few green- or brown-stained pebbles of limestone occur at the extreme base and were presumably derived from the underlying pebble-bed. (About 2 feet.)

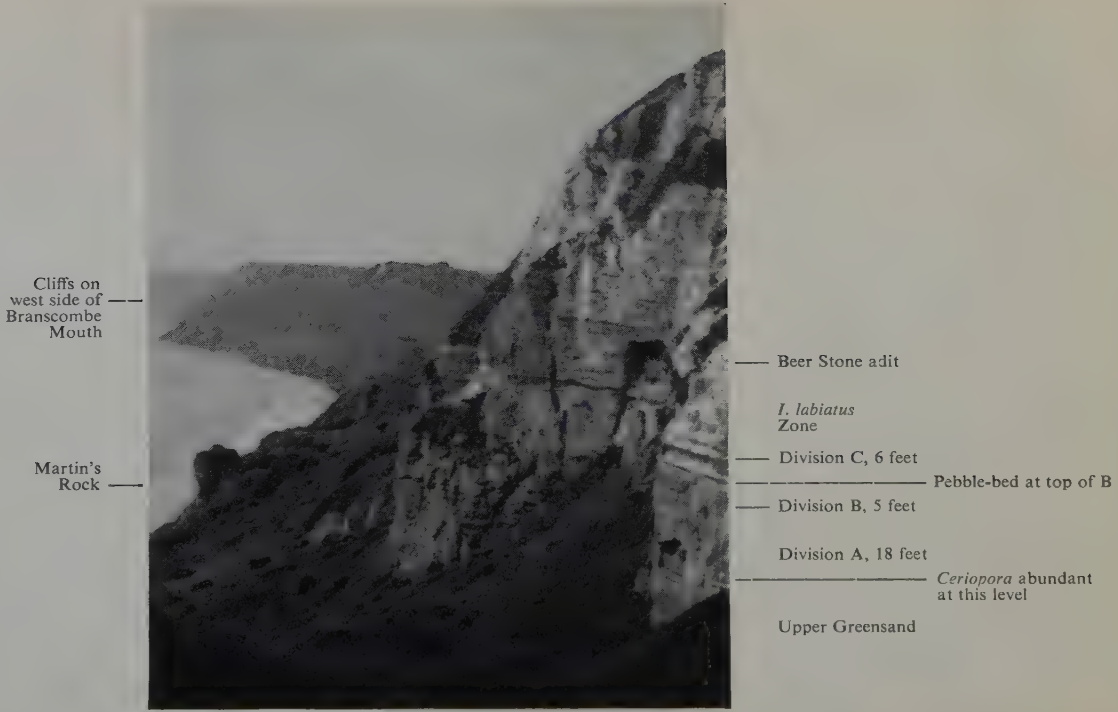
Division A. Thin impersistent pebble-bed containing small ellipsoidal green- and brown-stained pebbles of limestone. Top surface of pebble-bed or, when this is absent, the top surface of the underlying limestone is very even and appears to be a phosphatised erosion-surface. (0 to 1 inch.)

Nodular limestone containing rounded masses of hard white sandy limestones in a softer, sandier matrix. Little or no glauconite but brown and green



junction between the Middle Chalk and the Cenomanian deposits; westernmost nacle, Hooken Landslip.

he hammer rests on the erosion-surface forming the top of Division A of the omanian Limestone; the nodules and pebbles in the upper part of Division B have n rendered conspicuous by atmospheric weathering. Division C has weathered as cess between the nodular *Inoceramus labiatus* chalk and the pebble-bed at the top Division B



The eastern end of Hooken Cliff viewed westwards from the top of the footpath through the undercliff. The conspicuous recess in the projecting part of the cliff east of the Beer Stone adit indicates the position of Division C of the Cenomanian Limestone. The total thickness of Cenomanian deposits in this section is about thirty feet but the beds below Division C are obscured by chalk-wash. The adit indicates streaks are frequent near the top and appear to be the stained exteriors of blocks of limestone.

Aequipecten asper (Lamarck), *Cyclothyris* cf. *dimidiata* (J. de C. Sowerby), *Discoidea subuculus* Klein and *Holaster laevis* de Luc. (Up to 9 feet seen.)

(b) The Pinnacles and Other Displaced Blocks in the Hooken Landslip

The clearest section is exposed on the weathered northern face of the
ore westerly Pinnacle. The visible thickness of Cenomanian deposits is
re about twenty-five feet but since the lower part of Division A is
ncealed by talus and vegetation, the total thickness must be slightly
eater. The Pinnacle has come to rest with its component strata lying
most horizontally, so that the upper beds are not readily accessible. It is
ear, however, that the sequence is closely similar to that in the foundered
ock on the beach below (Plate 4). Division C is, however, distinctly
ndier than the equivalent bed in the block on the beach, a feature which
ay to some extent be the result of partial decalcification. Fossils collected
om Division C at this locality include a weathered belemnite-guard,
ctinocamax cf. *plenus* (Blainville), and the ammonites, *Schloenbachia* ? sp.
ov. aff. *lymense* Spath and *Calycoceras* sp.

An interesting section is provided by a small block lying in a clearing
out one hundred yards north-west of the Pinnacles and south of the main
otpath through the landslip, opposite the Beer Stone adit (30/219879).
he lowest bed seen in this block is a firmly cemented accumulation of
een- and brown-stained pebbles of hard limestone and is evidently that
hich forms the top of Division B. It is, however, considerably thicker than
the more easterly sections, being at least one foot thick. The pebble-bed
overlain by seven feet of Division C, the thickest accessible development
this bed in the Hooken Landslip. *Orbirhynchia wiesti* (Quenstedt) with
phosphatised valves filled with chalky limestone is abundant throughout
e bed. Poorly preserved phosphatic internal casts of other rhynchonellids,
cluding wide forms reminiscent of cyclothyrids, are not uncommon,
pecially in the lower part of the bed. Indeterminate phosphatised abraded
agments of ammonite casts also occur near the base. Jukes-Browne (1903)
cords a derived phosphatic cast of *Acanthoceras hippocastanum* from
ivision C in the Hooken Landslip. If correctly determined this is a
rticularly important record since the species does not occur either in
ivision A or Division B of the Cenomanian Limestone and must have
en derived from a bed which has been completely removed from the
er district by penecontemporaneous erosion. It would therefore appear
at, not only does Division C contain indigenous fossils and derived and
manié Cenomanian forms of several different ages, but also that the
ratigraphical break between this bed and Division B must mask a
mplex sequence of alternating periods of deposition and erosion.

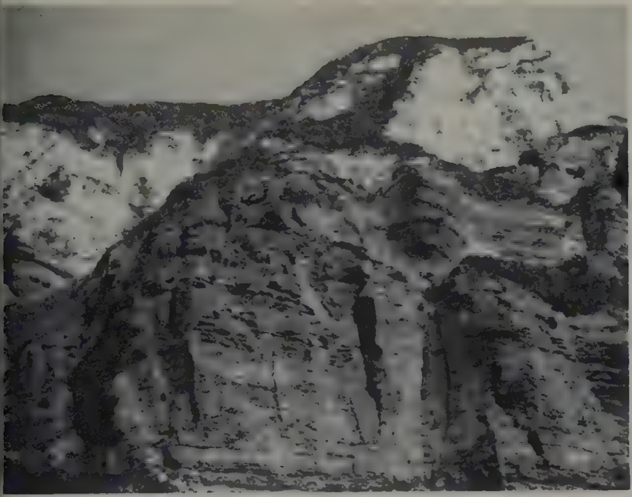
(c) The Eastern End of Hooken Cliff (30/220881)

The section immediately east of the Beer Stone adit in Hooken Cliff is very important since it is the only section displaying the entire thickness of the Cenomanian deposits at this end of the landslip. Division C is about six feet thick and has weathered to form a conspicuous recess beneath the harder nodular chalk of the *Inoceramus labiatus* Zone (Plate 5). Division B is about five feet thick and closely resembles that seen in the Pinnacles. Division A is badly obscured by chalk-wash but the junction with the Upper Greensand is visible here and there. The bed is about eighteen feet thick and, as in the section at Beer Head (Smith, 1957a), the base rests upon an erosion-plane forming the upper surface of the calcareous sandstone at the top of the Upper Greensand. The basal part also resembles that at Beer Head in being a coarse-grained calcareous grit containing numerous *Ceriopora ramulosa* Michelin. If Division C be classed as Cenomanian then the total thickness of Cenomanian deposits in the section is nearly thirty feet, a thickness which is greater than at any other locality in the cliffs of south-east Devon.

(d) The Stratigraphical Age of Division C

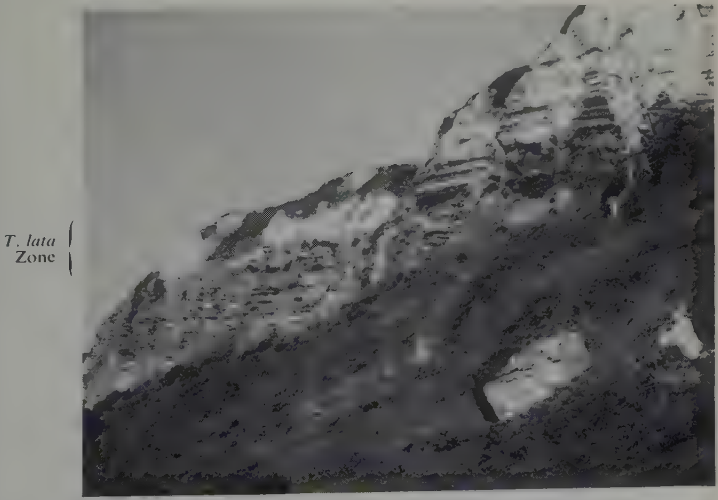
With the exception of certain long-ranged species such as *Discoidea subuculus* Klein, *Conulus castanea* (Brongniart) and *Cyclothyris* cf. *dimidiata* (J. de C. Sowerby) the indigenous fauna of Division C is quite distinct from that of Division B. The stratigraphical break between the two divisions is undoubtedly of major importance so that Division C must be considerably younger than the Lower Cenomanian limestone upon which it rests.

All the ammonites so far collected by the author from the exposures of Division C in the Hooken Landslip are interpreted as derived or *remanié* forms. They do not establish a Cenomanian age for the deposit but indicate the reworking of strata of that age. The Plenus Marls have until recently been grouped with the Cenomanian but their ammonite fauna suggests that they should be classed as a subzone of the Turonian *Inoceramus labiatus* Zone (Spath, 1926, 425; Wright & Wright, 1951, 4). The index-fossil of the *Actinocamax plenus* Subzone, however, ranges down into the top of the *Holaster subglobosus* Zone so that a correlation between Division C and the Plenus Marls cannot be regarded as firmly established on the belemnite evidence alone. *Orbirhynchia wiesti* (Quenstedt) is very closely related to, and sometimes almost indistinguishable from, *O. cuvieri* (d'Orbigny), the species which has been used instead of *Inoceramus labiatus* as an index-fossil for the lower zone of the Middle Chalk. *O. wiesti* (d'Orbigny) has been recorded as a rare fossil in Division B (Meyer, 1874; Jukes-Browne & Hill, 1896) but the present author doubts whether the species, as described



. Hooken Cliff, looking northward from a point on the footpath through the under-
iff, about 250 yards east of Martin's Rock.

The nodular chalk of the *Inoceramus labiatus* Zone (which has weathered with a
uted surface) and the underlying Cenomanian Limestone are both considerably
inner than in the vicinity of the Beer Stone adit to the east, Plate 2



. The western end of Hooken Cliff viewed from a point on the footpath through the
ndercliff, about 250 yards east of Martin's Rock.

The *Inoceramus labiatus* Zone of the Middle Chalk is not present in the cliff on the
ft-hand side of the photograph; the *Terebratulina lata* Zone resting directly upon a
reatly attenuated development of the Cenomanian Limestone, see Fig. 4b

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/ Pettitt (1954), ranges down into that division. The precise stratigraphical nge of *Orbirhynchia wiesti* (Quenstedt) is at present uncertain but Pettitt 954, 38) regards this species as occurring both in the upper part of the enomanian and in the *Actinocamax plenus* Subzone. The evidence at esent available is therefore insufficient to date Division C precisely. The posit could be of late Cenomanian age, early Turonian age, or could nge from late Cenomanian to early Turonian.

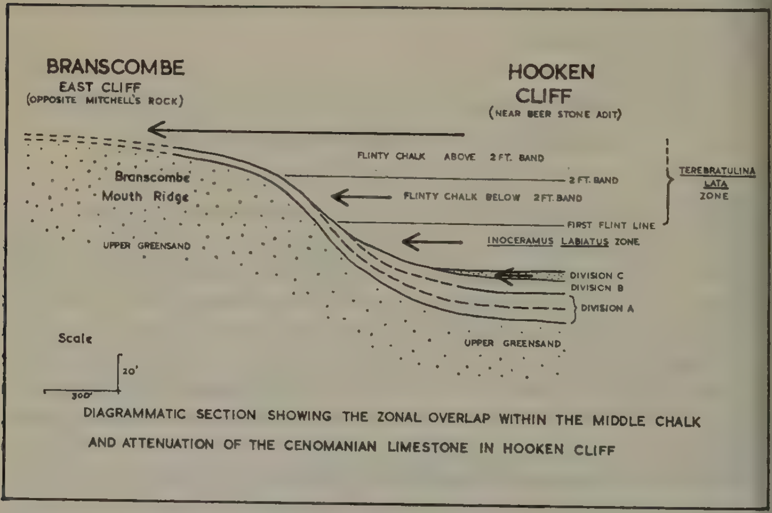
For the purpose of the present paper Division C is regarded as being of te Cenomanian age but it is clear that further work, particularly the scovery of undoubted indigenous ammonites, may show that the whole part of the deposit should be classed as being of early Turonian age. ıch a step would necessitate a new name for the deposit, since it could no nger be classed as a division of the Cenomanian Limestone, but would in keeping with the fact that Division C is physically a local sandy ısement-bed of the Middle Chalk. In order to avoid possible confusion a later date the name *Orbirhynchia* Band is therefore proposed as an ternative to Division C.

3. SECTIONS AT THE WESTERN END OF THE HOOKEN LANDSLIP AND ADJACENT LOCALITIES

(a) The Western End of the Hooken Landslip

West of the Beer Stone adit the Cenomanian rises in the steep vertical ıce of Hooken Cliff and is quite inaccessible not only because of the eepness of the cliff but also because of the dense vegetation between the ɔotpath and the cliff-face. It is quite clear, however, that the thick evelopment of the Cenomanian at the eastern end of Hooken Cliff is not ɔntinued westwards. The most conspicuous features of the section at the estern end of Hooken Cliff is the remarkable westerly overlap which ccurs within the Turonian (Whitaker, 1871; Jukes-Browne, 1903; Rowe, 903). The nodular chalk of the *Inoceramus labiatus* Zone which is so ɔnspicuous in Hooken Cliff wedges out westwards since each bed of chalk overlapped in this direction by the bed overlying it. Towards the western ıd of Hooken Cliff the *Inoceramus labiatus* Zone is absent since it is ɔmpletely overlapped by the overlying flinty *Terebratulina lata* Zone ?late 6 A, B). The lower beds of the latter zone, including the bed of marly intless chalk termed the '2 ft. band' by Rowe, are also eventually overıpped by higher beds. At the western end of the section (i.e. in Branscombe ast Cliff) flinty chalk representing an horizon well above the base of the *erebratulina lata* Zone rests directly upon a projecting ledge of Cenoıanian Limestone which is about three feet thick. It is possible that at the xtreme western end of Branscombe East Cliff the *Terebratulina lata* Zone omes to rest directly upon the Upper Greensand, but the section is not sufficiently clear to permit confirmation of this. Rowe has estimated that about seventy-three feet of chalk are overlapped between the Beer Stone adit and Mitchell's Rock, a distance of about one-third of a mile. In the same distance at least twenty-seven feet of Cenomanian deposits disappear so that about one hundred feet of strata present in the eastern end of Hooken Cliff are not represented in Branscombe East Cliff.

The inaccessible nature of the cliff-section makes it difficult to determine the lateral changes in the Cenomanian. An attempt was made to follow various horizons along the cliffs with the aid of high-power binoculars but was only partly successful because of deep decalcification and carious weathering of both the Cenomanian Limestone and the underlying Upper Greensand. The evidence obtained from this visual examination of the cliff was supplemented by the study of fallen blocks in the undergrowth north of the footpath through the landslip.



DIAGRAMMATIC SECTION SHOWING THE ZONAL OVERLAP WITHIN THE MIDDLE CHALK AND ATTENUATION OF THE CENOMANIAN LIMESTONE IN HOOKEN CLIFF

Fig. 4 b

The conspicuous recess formed by the soft sandy glauconitic Division C persists for only a short distance westwards and the deposit appears to be overlapped by the nodular *Inoceramus labiatus* chalk. Division C is therefore not only the local basement-bed of the Middle Chalk but it also appears to be the earliest member of a transgressive group of rocks (Fig. 4b). It is not clear, however, whether the thinning of Divisions A and B of the underlying Cenomanian Limestone is due to overlap or is due to overstep by the transgressive Middle Chalk.

few yards east of the fallen 'Mitchell's Rock' (30/214881), there is a spicuous block, tilted landwards and exposing over twenty feet of per Greensand, capped by thirty-three inches of coarse pebbly grit taining small ill-preserved fragments of *Ceriopora* and many shell-ments, especially those of *Pecten*. The lithology and its abrupt base on gularly decalcified Upper Greensand suggest that this grit can be related with the coarse basal part of Division A near the Beer Stone adit. Underneath the ivy on the northern side of Mitchell's Rock there are y inches of crumbly, partially decalcified grit between the Upper ensand and the soft, white, flinty chalk of the *Terebratulina lata* Zone. sils are not common in the grit, except at the very top, where polyzoan ments and *Cyclothyris dimidiata* (J. de C. Sowerby) are not uncommon. es-Browne (1903, 138) has published a more complete list of fossils ected from this exposure. In the inaccessible cliff-section north of chell's Rock the projecting ledge beneath the *Terebratulina lata* chalk ears to be the same grit. There are no fossils indicative of higher levels he Cenomanian Limestone in this grit so that it is extremely unlikely t it is a condensed sequence equivalent to the whole of that formation. absence of Division B and the upper part of Division A shows that thinning of beds earlier than Division C to the west of the Beer Stone t must be due to overstep by the Middle Chalk (Fig. 4a). The uncon-nity indicated by this overstep can be clearly seen when the entire oken section is viewed from the sea; the easterly inclination of the ddle Chalk can then be seen to be appreciably smaller than that of the lerlying strata.

he erosion of Cenomanian strata implied by the westerly overstep rds a reasonable explanation of the abundance of sand and glauconite the presence of a varied derived fauna in Division C. Nevertheless, a y surprising feature of the Hooken Cliff section is the absence of any ious lithological change in the various Middle Chalk beds as they cessively come into contact with sandy Cenomanian deposits. The lular beds in the *Inoceramus labiatus* Zone and marly beds in the *ebratulina lata* Zone appear to retain their distinctive lithological racters right up to the point at which they are overlapped. The chalk nediately above the Cenomanian Limestone in Mitchell's Rock is white marly but contains only traces of sand despite the fact that it rests ectly upon coarse-grained Cenomanian calcareous sandstone. The early nomanian sandy deposits, and possibly the upper part of the Upper eensand near Branscombe Mouth, were clearly within reach of wave-ack or current-scour during the accumulation of Division C farther east. entually, however, following subsidence or a slackening of the marine rents, the intensity of the erosion gradually decreased and during the umulation of the *Inoceramus labiatus* Zone and the lower part of the

Terebratulina lata Zone chalk containing very little sand was deposited against a slope forming the eastern side of a submarine ridge on the early Turonian sea-floor in the Branscombe Mouth region. Marine currents passing over this ridge must have been sufficiently strong to prevent the accumulation of chalk-mud but, after the accumulation of Division C, were not strong enough to produce any appreciable erosion of the sandstones laid bare on the sea-floor. The area of this region of inhibited deposition gradually decreased as the ridge was slowly buried beneath chalk-mud until beds well above the base of the *Terebratulina lata* Zone were laid down in direct continuity with those accumulating west of Branscombe Mouth.

(b) Stockham's Hill Quarry

An important exposure of the Cenomanian Limestone occurs in the partially overgrown quarry on the western flank of Stockham's Hill, about half a mile north of Branscombe Mouth and north-west of Hooken Cliff (30/207888). The quarry-face extends in a general north–south direction and displays a section which differs considerably from those seen in Mitchell's Rock and in the adjacent cliff-face. Throughout its entire thickness the Cenomanian Limestone in the quarry is a hard white glauconitic limestone yielding ammonites such as *Schloenbachia subtuberculata* (Sharpe) and is clearly a development of Division B. The limestone rests abruptly upon tufa-encrusted calcareous grit, about ten feet of which is exposed in the deepest part of the quarry. This grit contains numerous small quartz pebbles, especially in the upper part, but passes down without a break into finer-grained calcareous sandstones lithologically identical with those near the top of the Upper Greensand at Hooken Cliff. Although scarce ill-preserved rhynchonellids occur in the grit, fossils which are common in the coarse-grained basal part of the Cenomanian Limestone, e.g. large polyzoa and cyclothyrids, are completely absent. The pebbly grit is clearly a particularly coarse-grained development of the top of the Upper Greensand so that there would appear at first sight to be nothing in the section equivalent to Division A of the Cenomanian Limestone. However, on a large fallen block in the quarry, Jukes-Browne (1903, 138) noticed a very thin bed containing numerous pebbles of sandstone and *Trigonia* shells and he regarded this bed as providing evidence that Division A had been deposited in the area and was removed subsequently by intra-Cenomanian erosion.

The surface of the Upper Greensand beneath the Cenomanian Limestone is uneven, depressions several feet in width alternating with low elevations elongated in an east–west direction, roughly normal to the quarry-face. In contrast the top surface of the Cenomanian Limestone does not undulate but is a plane inclined gently towards the east as a result of the tectonic

f the Cretaceous strata. The Cenomanian Limestone therefore displays
d lateral variations in thickness being up to twenty-seven inches thick
the depressions in the surface of the Upper Greensand but only eight
ine inches thick over the intervening elevations. The depressions in the
ace of the Upper Greensand appear to be shallow channels produced
urrent-scour. By analogy with adjacent areas the top of the Upper
ensand was probably eroded by marine currents before the accumula-
of any deposits equivalent to Division A. Nevertheless, the surface of
Upper Greensand as seen today must have been produced after the
oval of these early Cenomanian deposits, during a subsequent erosion-
od which preceded the accumulation of Division B.

he Cenomanian Limestone in Stockham's Hill quarry is overlain by
ut five feet of flinty chalk-wash and 'clay with flints', but in one or two
es patches of undisturbed chalk occur in the lower part of the over-
den and appear to be *in situ*. The chalk is hard and compact immediately
ve the Cenomanian Limestone but otherwise is mostly soft, thin-
ded and contains small horned flints. No macrofossils have so far been
d in this chalk, its zonal position is therefore doubtful but the general
ology suggests that it forms part of the *Terebratulina lata* Zone.

would appear from the foregoing description that the zonal overlap
in the western end of Hooken Cliff persists at least as far north as
kham's Hill. The excavation of the Branscombe Valley has, however,
pletely destroyed the evidence needed to determine the direction of the
rlap at the latter locality. It could even be in the opposite direction to
westerly overlap seen in Hooken Cliff. Since the *Inoceramus labiatus*
e is well developed west of Branscombe Mouth, the absence of this
e in Stockham's Hill quarry could be the result of easterly overlap
inst the western flank of the submarine ridge which must have existed
he Branscombe Mouth region during early Turonian times. In either
it is clear that Stockham's Hill quarry must lie near the crest of this
ge.

4. SECTIONS WEST OF BRANSCOMBE MOUTH

(a) Branscombe West Cliff to Donkey Linhay

he upper surface of the Cenomanian Limestone, overlain by remnants
he basal Middle Chalk, is laid bare for a short distance at the top of the
tern end of Branscombe West Cliff. Towards the west the formation
ses beneath the precipitous outcrop of the *Inoceramus labiatus* Zone and
the remainder of Branscombe West Cliff and in Berry Cliff is largely
cealed beneath dense vegetation. Most of these exposures are quite
ccessible but, since the Cenomanian Limestone often forms the top of a
jecting ledge below the outcrop of the Middle Chalk, it is possible to
n some idea of lateral variations in the thickness of the various Ceno-

manian beds by viewing sections in small bluffs across narrow embayments in the cliff-line. At the western end of Berry Cliff, above Donkey Linhay Rocks, the Cenomanian Limestone immediately underlies a well-defined footpath running along the ledge below the Middle Chalk and is clearly exposed for nearly 200 yards. Close study of this section is, however, rendered difficult by the precipitous nature of the cliff below the footpath. Between the western end of the Donkey Linhay footpath and the small valley known as 'Littlecombe Hollow' there is no clear section because of the dense vegetation which covers the entire face of the cliff.

For a detailed study of the lithology and fauna it is necessary to study fallen blocks in the undercliff and on the beach. Such blocks are smaller and less numerous than in the Hooken Landslip but are not uncommon among the rocks forming Branscombe Ebb, about three-quarters of a mile west of Branscombe Mouth. Most of these blocks lie below the high-water mark and are encrusted with marine growths, but those at the rear of the beach at the eastern end of the Ebb display clean sections exposing the entire thickness of the Cenomanian Limestone.

(*i*) *The Cliff-Sections*. Despite the intermittent nature of the exposures it is clear that the top surface of the Upper Greensand is uneven due to the presence of erosion-channels similar to those seen at Stockham's Hill quarry. As in the latter section the erosion-channels appear to be elongated in an E.–W. direction since southerly thinning of the lower part of the Cenomanian Limestone, presumably towards the southern side of a channel, can be seen in some of the bluffs, especially those above The Cove, about half a mile west of the old coastguard station at Branscombe Mouth.

The lowest bed in the Cenomanian Limestone in Branscombe West Cliff consists of coarse calcareous grit which resembles the basal part of Division A as developed in Hooken Cliff. This coarse calcareous grit, termed 'Division A 1' by Jukes-Browne, varies greatly in thickness, being several feet thick in places and only a few inches or completely absent in others. The rapid variation in thickness appears to be largely due to the deposition of the calcareous grit upon the undulating erosion-surface at the top of the Upper Greensand.

Division A 1 is overlain by about two feet of very hard, shelly sandy limestone which resembles the upper part of Division A as developed farther east and termed 'Division A 2' by Jukes-Browne. The bed appears to be persistent in the cliffs west of Branscombe Mouth and rests directly upon the Upper Greensand in places where Division A 1 is absent. Despite its persistence it is not of uniform thickness. It appears to be thinnest where the underlying Division A 1 is particularly thick. For example, in one bluff above The Cove, Division A 1 is about six feet thick and Division A 2 only nine inches whereas about 100 yards farther west Division A 1 is absent and Division A 2 nearly three feet thick. It is therefore probable that

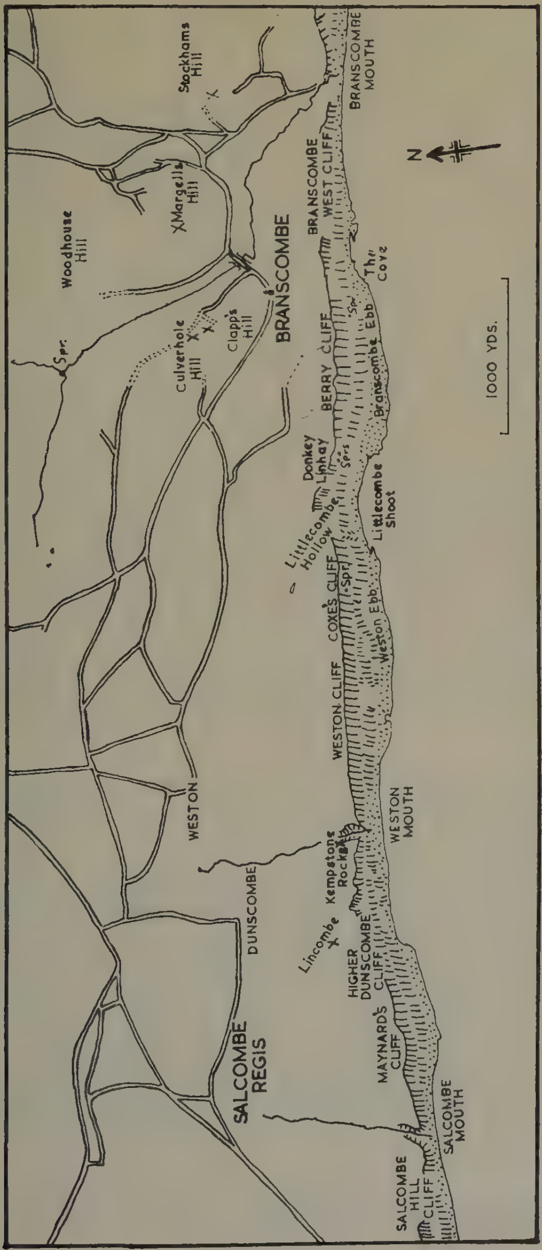


Fig. 5. Locality-map of the area between Branscombe and Salcombe Regis, SE. Devon

the lateral variation in the thickness of Division A1 is due not only to deposition on an irregular surface but is also due to the irregular erosion of the coarse calcareous grit before the deposition of Division A2. This is particularly significant since the writer has suggested that an erosion-period which preceded the deposition of shelly Cenomanian Limestone removed coarse Cenomanian grit from the Beer district (Smith, 1957a).

Division B of the Cenomanian Limestone appears to be represented by a few inches of sandy limestone containing pockets of glauconitic sand and green- and brown-stained phosphatised pebbles of limestone. This bed is quite distinct from the underlying white shelly limestone and is overlain abruptly by the yellowish basement-bed of the Middle Chalk. As elsewhere, this basement-bed contains scattered grains of quartz and glauconite but never develops the lithology characteristic of Division C.

Under the lower cliff-path at the western end of Berry Cliff, above Donkey Linhay Rocks, both Division A1 and Division B appear to be absent. The basement-bed of the Middle Chalk is separated from the Upper Greensand by only two feet of Division A2 which is here a very hard white limestone containing scattered well-rounded grains of quartz, scarce granules of glauconite and occasional lamellibranch shells and shell-fragments. This very incomplete development of the Cenomanian Limestone persists at least as far as the western end of the footpath.

(*ii*) *Fallen Blocks on the Beach at Branscombe Ebb.* The fallen blocks at the rear of the beach at the eastern end of Branscombe Ebb provide several clean sections displaying thick developments of the Cenomanian Limestone. The largest and most conspicuous block displays the following section:

> **Basal Middle Chalk.** Hard white chalk containing very little sand or glauconite, junction with the underlying bed clear-cut but slightly irregular. (4–5 inches.)
>
> **Division B of the Cenomanian Limestone.** Nodular, green- and brown-stained limestone with pockets of glauconitic sand, rests upon a brown-stained phosphatised erosion-surface at the top of the underlying bed. (4–5 inches.)
>
> **Division A of the Cenomanian Limestone.** Hard white shelly limestone (A 2) passing down into sandy limestone containing abundant grains and small well-rounded pebbles of quartz. *Ceriopora ramulosa* Michelin is common at or near the base. The lower part is evidently a development of Division A 1 but there is no clearly defined boundary separating it from Division A 2. Rounded masses of sandy limestone, several inches in diameter, are common in the lower part; these are lithologically distinct from their host-rock and are presumably water-worn blocks of early Cenomanian Limestone washed in from an adjacent area during the accumulation of Division A 2. (Total thickness 6 feet.)

An unusual feature of the blocks at the eastern end of Branscombe Ebb is the presence of chert in the lower division of the Cenomanian Limestone.

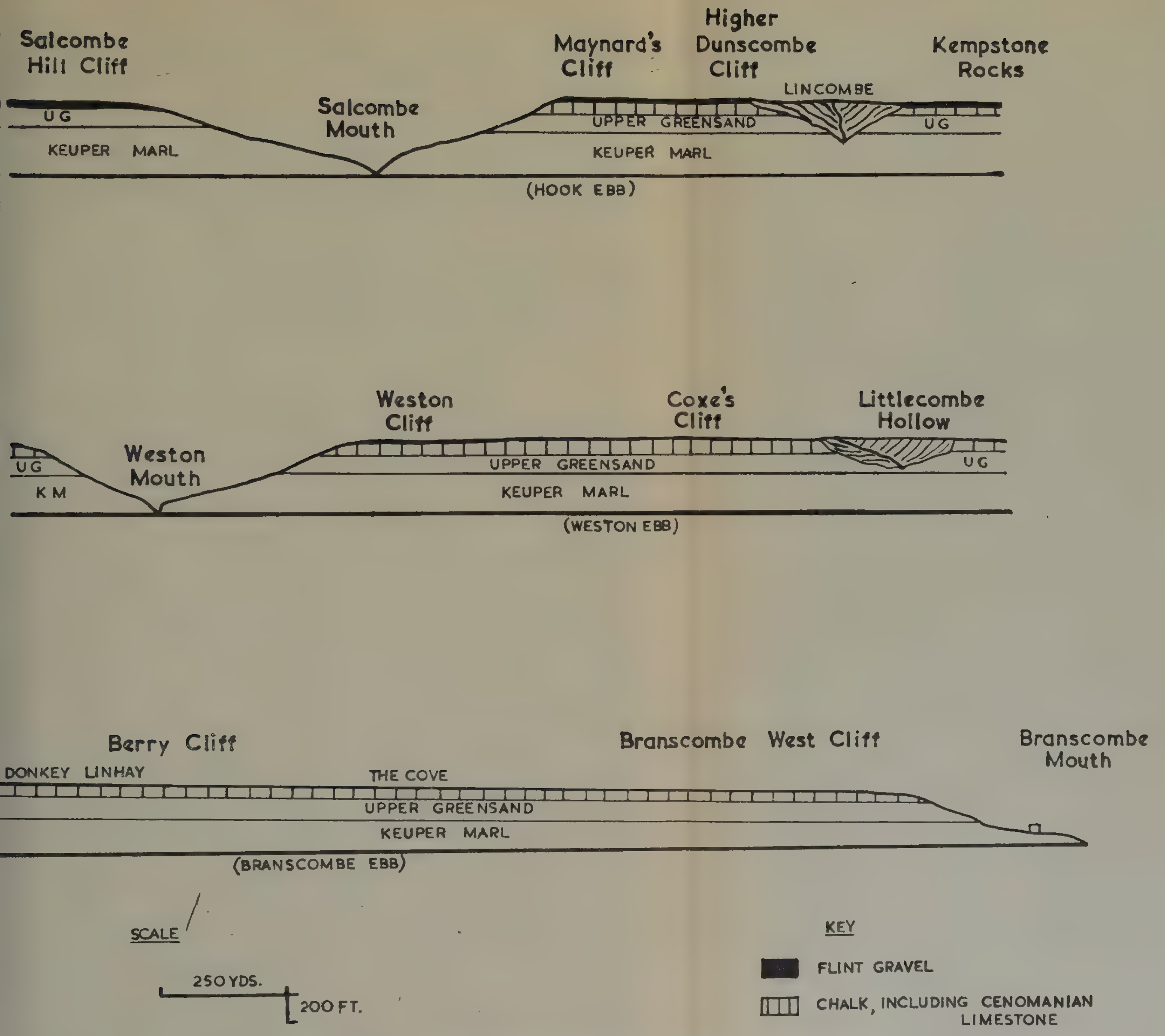


Fig. 6. The cliff-sections between Branscombe Mouth and Salcombe Hill Cliff, near Sidmouth

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r example, a small block of Division A1, lying close to the large block
scribed above, contains abundant large *Ceriopora* and a complex
erlacing system of irregularly shaped veins of greyish chert up to nine
ches in length. The chalcedonic matrix of the chert contains abundant
icified polyzoan zoecia; unsilicified zoecia of identical character also
cur in calcareous inclusions and embayments within the chert. The
mplex arrangement of the chert veins is consistent with the view that
ey form part of an incompletely silicified lamellar polyzoan possessing a
isted or convolute zoarium. Most of the insoluble residues left after the
eatment with dilute hydrochloric acid of samples of Division A collected
om all its major outcrops contain numerous silicified fossils, including
hall silicified polyzoa; the grey chert at Branscombe Ebb would appear
be a particularly large and conspicuous example of the latter.

(b) Littlecombe Hollow to Salcombe Mouth

(i) *Littlecombe Hollow to Weston Mouth.* In the cliffs between Little-
mbe Hollow and Weston Mouth, i.e. in Coxe's Cliff and Weston Cliff,
e Cenomanian Limestone rises to a height of nearly 500 feet and is quite
accessible. Fallen blocks under Weston Cliff show, however, that the
enomanian Limestone is more complete here than at Donkey Linhay
id, as at Branscombe West Cliff, consists of three beds. The two lower
ds resemble those comprising Division A in Branscombe West Cliff. The
isal bed is again a coarse-grained calcareous grit containing large
eriopora and many lamellibranch shells, its thickness varying from place
place but averaging about two feet. The overlying bed is of more
niform thickness and consists of about two feet of very hard, compact
elly limestone containing scattered grains of quartz but little glauconite.
his bed is clearly Division A2 but is overlain by about one foot of hard
hite limestone containing larger and more numerous grains of quartz but
ss glauconite than are present in Division B in the more easterly sections.
he presence of *Holaster subglobosus* Leske and *Cyclothyris dimidiata*
. de C. Sowerby) and the complete absence of *Orbirhynchia wiesti*
uenstedt) show, however, that despite its distinctive lithology, the bed
ust be regarded as a development of Division B and must not be confused
ith the sandy limestone in Hooken Cliff referred to as 'Division C'. This
nclusion is supported by the fact that the hard sandy limestone at
eston Cliff is overlain abruptly by the basal Middle Chalk which, as
sewhere, consists of hard cream-coloured chalk containing only scattered
ains of quartz and glauconite.

(ii) *Weston Mouth to Salcombe Mouth.* In this area the Cenomanian
imestone can only be examined *in situ* at the western end of a small
isused quarry in the top of the cliff at the eastern end of Kempstone
ocks (30/161881). Three divisions can be recognised, the lowest of which,

Division A1, is about two feet thick at the rear of the section but thins out completely towards the southern end of the quarry-face. The bed appears to rest in a scour-channel cut in the surface of the Upper Greensand and trending E.–W., i.e. in a similar direction to the scour-channels in the top of the Upper Greensand in the Branscombe cliffs and at Stockham's Hill. Division A2 is a hard compact shelly limestone and maintains a constant thickness of about two feet; it rests directly upon the Upper Greensand at the southern end of the section. Division B is represented by about eighteen inches of whitish gritty limestone similar to that in Weston Cliff.

In Higher Dunscombe Cliff and Maynard's Cliff, east of Salcombe Mouth, the Cenomanian Limestone is again inaccessible and largely concealed by gravel-fans and vegetation. The limited exposures suggest, however, that the three divisions seen in Kempstone Rocks are continuous throughout these cliffs and show a general westerly increase in thickness. The chalk bluff at the western end of Maynard's Cliff, the westernmost coastal outcrop of chalk in southern England, shows that the total thickness of the Cenomanian Limestone is here about eight or nine feet. This is confirmed by the fallen blocks on the beach, especially those at the western end of the Hook Ebb rocks, about half a mile east of Salcombe Mouth. These blocks have been considerably eroded by wave-action since the days of Jukes-Browne and are encrusted by marine growths, but the main lithological details can still be made out, particularly in one very large inverted block. Wave-smoothed blocks, partially buried in the shingle, also provide numerous sections in which the lithological features of parts of the succession are very clearly displayed. The succession in the large inverted block can be summarised as follows:

Division B of the Cenomanian Limestone

3. Hard white sandy limestone containing numerous large quartz grains and small inconspicuous granules of glauconite, green-stained masses of sandy limestone scattered throughout the bed. *Holaster subglobosus* Leske. (2 feet.)

—EROSION-SURFACE—

Division A 2 of the Cenomanian Limestone

2. Hard white sandy limestone containing little or no glauconite, basal part contains current-rounded *Ceriopora* and rounded masses of pebbly grit and sandy limestone some of which are encrusted with polyzoa. (4 feet.)

Division A 1 of the Cenomanian Limestone

1. Calcareous grit containing numerous large sand-grains and pebbles of clear, milky and opaque white quartz, many having diameters exceeding 5 mm. No sharp junction with the overlying bed but the frequency and maximum grain-size of the enclosed quartz grains decreases rather abruptly at the presumed junction. *Ceriopora ramulosa* Michelin in white chalky limestone common, especially towards base. (2 feet.)

The rounded masses of pebbly grit and limestone at the base of Division A2 are particularly conspicuous in some of the small beach-blocks lying n the shingle. Their lithology, together with the abraded nature of the *Ceriopora*, clearly indicate a period when the lower part of Division A was ndergoing erosion in an adjacent area. It is probable that there was no mportant break during the deposition of Division A in the Maynard's Cliff area but that during the erosion-period suggested by the erosion-lane which separates Divisions A1 and A2 farther east, blocks of lime-tone and calcareous grit were washed into this area. These blocks lay on he sea-floor for a sufficiently long period to allow some of them to become ncrusted with polyzoa but there was no real halt in deposition. It is very ifficult to locate the region from which the rounded blocks were derived. The complete absence of Division A1 in the Donkey Linhay cliffs suggests hat the blocks could have been derived from that area. On the other hand he removal of the Cenomanian Limestone west of Salcombe Mouth by ost-Cretaceous erosion has destroyed any evidence of the penecontem-oraneous erosion of Cenomanian deposits in the Sidmouth area.

The sandy development of Division B must maintain a fairly constant ithology between Maynard's Cliff and Weston Cliff. It is thickest in Maynard's Cliff, decreases in thickness towards the east and thins out ltogether in the Donkey Linhay area. East of Donkey Linhay Division B eappears in Berry Cliff as a thin bed containing more glauconite but less and and tends to increase in thickness towards the east, i.e. towards Branscombe West Cliff (Fig. 8).

(c) The Western Limit of the Cenomanian Limestone

The thickness of the Cenomanian Limestone in the fallen blocks under Maynard's Cliff differs slightly from those recorded by Jukes-Browne. The nain difference is in the thickness of Division A1 which Jukes-Browne ives as about four feet, i.e. about twice the thickness recorded above. The iscrepancy is presumably due to marine erosion of the beach-blocks uring the past fifty years. If Jukes-Browne's figures be accepted as orrect, the total thickness of Cenomanian Limestone in its westernmost utcrop is about ten feet, i.e. greater than at any other locality along the outh Devon coast except Beer Head and Hooken Cliff. It is not possible, owever, to determine whether this comparatively great thickness of Cenomanian Limestone in Maynard's Cliff is merely a local feature since he formation is completely absent west of Salcombe Mouth. There is, owever, no evidence to suggest that the Cenomanian was not laid down arther west, although the very coarse nature of the earliest Cenomanian eposits suggests that, at least during early Cenomanian times, the shore-ne was not far distant. The vertical change in the lithology, particularly he decrease in the size of the enclosed sand-grains, is consistent with the postulation of a Cenomanian transgression and retreat of shore-lines, although the decrease in size and abundance of detrital material may also have been due to a reduction in the relief of a westerly land-mass.

5. INLAND EXPOSURES OF CENOMANIAN ROCKS WEST OF SEATON

A guide to the position of any outcrop of the Cenomanian Limestone is provided by the greyish scars which project from the hillside turf along the outcrop of the top beds of the Upper Greensand. Exposures are, however, very rare because of the flinty or cherty downwash which mantles the upper flanks of the Chalk-capped hills. Quarry-sections and such natural exposures as do occur show that, except in certain areas, the formation is only one or two feet thick so that the width of its outcrop must be extremely narrow and not marked by any prominent feature. The exposures which have been found are all in the southern margins of the chalk-outliers because of the general south-easterly dip of the Cretaceous rocks.

(a) The Beer District

There are several exposures of the Cenomanian Limestone in and around Beer village. (1) About two feet of hard limestone comprising about equal thicknesses of Division A and Division B can be seen under the hedge at the top of the road-cutting opposite the Pioneer Garage. (2) Because of an easterly dip of about 10° the formation rises in the hillside which overlooks the Branscombe Road and can be seen above Gibb's Farm. (3) Still farther westwards the Cenomanian Limestone crops out in the shallow drainage-channel by the side of Quarry Lane about 200 yards east of Quarry Cottages. The exposure is very poor but the junction with the Upper Greensand can be laid bare. This junction is as abrupt as on Beer Beach, the hard white limestone resting directly upon calcareous sandstone with no evidence of the coarse calcareous grit which forms the base of the Cenomanian at Hooken Cliff, about one mile to the south. (4) Still farther west the top of the Cenomanian Limestone can be seen in the floor of the westernmost quarry south of Quarry Lane and opposite the main working quarry in the Middle Chalk (Fig. 2). The limestone has a similar lithology to that farther east, its upper surface being a clear-cut erosion-surface which is slightly phosphatised and has a gentle north-easterly slope. The junction with the overlying Middle Chalk is concealed beneath talus and piles of flints extracted from the working quarry on the opposite side of the road. Excavation failed to reveal any deposit similar to Division C as developed in Hooken Cliff. The lowest bed of the Middle Chalk is soft and contains very little glauconite and only a few sand-grains and small well-rounded pebbles of quartz. *Conulus castanea* (Brongniart) is common in the bed but no specimen of *Orbirhynchia wiesti* (Quenstedt) has so far been found in it. Apart from its soft nature the basal Middle Chalk in this quarry resembles that in the cliffs around Beer Beach; it is possible that the hardness of the basement-bed at the latter locality is due to its prolonged exposure or the action of sea-water. Despite the poor exposure of the Cenomanian Limestone west of Beer village it is clear that the great thickness of the Cenomanian strata at the eastern end of Hooken Cliff is not maintained towards the north.

The westernmost exposure in the Beer quarries is about half a mile north-east of the quarry on Stockham's Hill in which Middle Chalk, presumably of the *Terebratulina lata* Zone, rests upon the Cenomanian Limestone. The Middle Chalk above the Cenomanian Limestone in the Beer Stone quarries and in the many overgrown quarries on each side of Bovey Lane, about one-third of a mile to the north, includes an important development of the *Inoceramus labiatus* Zone which is thicker here than in Hooken Cliff. In other words, the northerly thinning of the Cenomanian deposits in Hooken Cliff is not accompanied by overlap within the Middle Chalk such as occurs towards Branscombe Mouth.

(b) The Bovey Lane Sand-Pit

One of the most surprising features of the inland exposures in the Beer district is that the Cenomanian deposits beneath the Middle Chalk north of the Beer Stone quarries are quite unlike those seen in the coastal sections and in the poor exposures along Quarry Lane. Cenomanian deposits, previously undescribed, are well exposed in a sand-pit under Bovey Coppice on the south side of Bovey Lane (30/217900, Fig. 2), and are at east twenty feet thick. The increase in thickness must be very sudden since he Bovey Lane sand-pit is only one-third of a mile north of Quarry Lane.

(*i*) *Section Exposed in the Bovey Lane Sand-Pit, Beer.*

6. Soft glauconitic sand with a chalky matrix and containing nodules of hard sandy glauconitic limestone, thin ochreous layer about two feet six inches above base.
Cyclothyris cf. *dimidiata* (J. de C. Sowerby), *Orbirhynchia wiesti* (Quenstedt), *Discoidea* sp. fragmentary ? *D. cylindrica* Lamarck.
A disturbed deposit, partly mixed with clay-with-flints and flinty hill-wash but near centre of pit thickness at least 5 feet.
A thin layer of soft brownish ochreous material resembling a weathered phosphatised crust.

5. Calcareous glauconitic sand passing up into white chalky limestone containing quartz and glauconite grains more or less uniformly distributed in the lower part but concentrated in streaks and lenticles towards the top, uppermost few inches distinctly nodular.
Rhynchonella grasiana d'Orbigny, *Discoidea subuculus* Klein, *Holaster laevis* de Luc, *Holaster subglobosus* Leske (abundant), *Pleurotomaria* sp. (internal cast).

Loose on floor of pit but assigned to this horizon by means of the lithology of the sediment associated with them:
Cyclothyris cf. *dimidiata* (J. de C. Sowerby), *Kingena arenosa* (d'Archiac), *Kingena* sp. (young undescribed form) *Echinogalerus* cf. *faba* (L. Agassiz), *Tetragramma variolare* (Brongniart), *Conulus castanea* (Brongniart) (6 feet.)

4. Calcareous sand consisting of quartz grains and very scarce glauconite contained in a white, friable, chalky matrix, numerous nodules or pebbles of hard white sandy limestone with long axes averaging about two to three inches. (3½ feet.)

3. Greyish calcareous sand, similar to 1, except in colour. (2 feet.)

2. Calcareous sandstone forming projecting ledge near base of quarry, similar to 1, except that the calcareous particles in the matrix are bonded together. (2¼ feet.)

1. Coarse-grained yellowish calcareous sands containing occasional more firmly cemented patches. Detrital minerals consist predominantly of quartz but also include pearly white and pinkish rectangular grains of orthoclase, some of which have a schiller reminiscent of murchisonite. Large well-rounded and prismatic grains of black tourmaline abundant and could be mistaken for particularly dark-coloured glauconite, but that mineral is either very scarce or completely absent. The various sand-grains are enclosed in a white, friable, chalk-like matrix. Base of deposit not exposed.
Concinnithyris sp., umbonal region; *Catopygus columbarius* (Lamarck), abundant; *Holaster laevis* de Luc, abundant; *Discoidea subuculus* Klein, *Aequipecten asper* (Lamarck), mainly fragmentary; *Pecten* spp., internal casts and abundant silicified shell-fragments; *Hallirhoa costata* Lamaroux, indeterminate sponge-remains. Thickness (in trial-hole) at least 10 feet.

Fossils loose on floor of pit but exact horizon uncertain:
Salenia austeni Forbes, probably from bed 1, 2 or 3.
Cottaldia bennettiae Koenig.
Mantelliceras sp., poorly preserved fragment probably from bed 5.
Nautilus sp. fragments of large species.

(*ii*) *Correlation of the Bovey Lane Section.* Despite a careful and prolonged search the poorly preserved *Mantelliceras* was the only ammonite found in the Bovey Lane sand-pit. In the absence of ammonites it is difficult to date the various beds precisely. The majority of the fossils which were found are comparatively long-ranged facies-fossils. The distinctive lithology and fauna of the very glauconitic bed at the top of the pit suggest that it can be correlated with Division C. The lithology and fauna of the underlying bed are closely similar to Division B in the Hooken Landslip (Fig. 7). The comparative abundance of *Holaster subglobosus* Leske in this bed is significant since this fossil is common in Division B in the coastal section but does not occur in Division A. *H. subglobosus* does not occur in the calcareous sands and sandstones in the lower part of the Bovey Lane sand-pit. These sands and sandstones are quite distinct from the sands at the top of the Upper Greensand. Their lithology and fauna agree closely with those of the Cenomanian sands at Wilmington, whilst their stratigraphical position in the Bovey Lane sand-pit suggests that they are a particularly sandy development of Division A (Fig. 7).

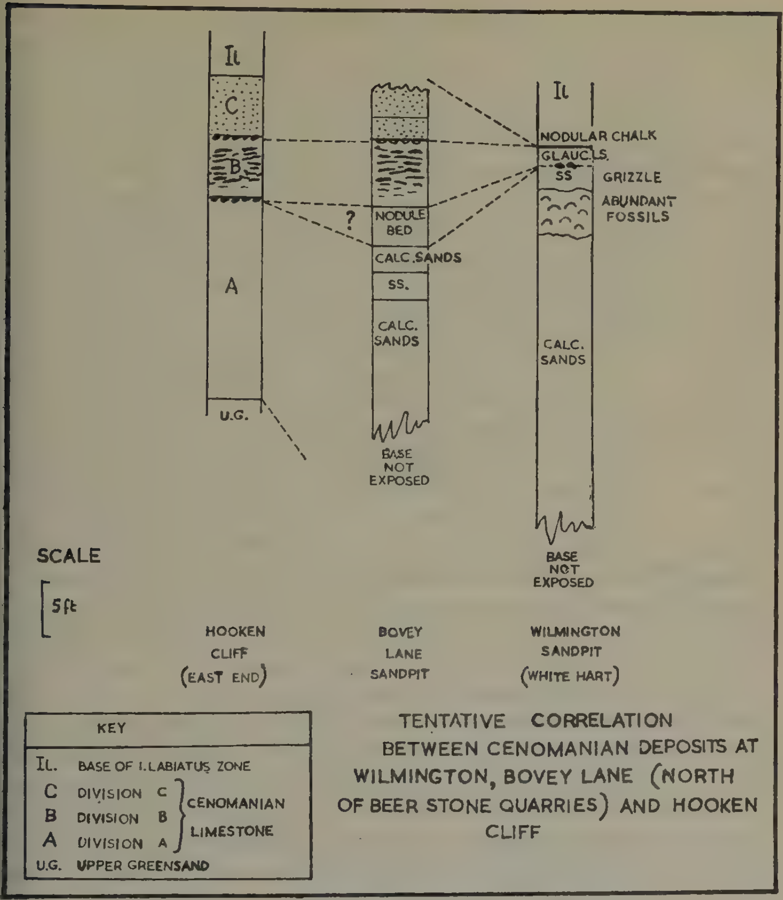


Fig. 7

The beds in the Bovey Lane sand-pit can therefore be classified as follows:

Very glauconitic bed at top of pit, at least 5 feet = Division C.
Sandy glauconitic limestone, about 6 feet = Division B.
Nodular bed, 3 feet 6 inches
Calcareous sand and sandstone, at least 15 feet } = Division A.

The lithology of bed 4 is unlike that of any other deposit in the Beer district. The nodules of hard sandy limestone resemble the incoherent matrix of the bed but their sand-content is distinctly lower. The insoluble sandy residue left after the treatment of five nodules with dilute hydrochloric acid indicates that their average sand-content is 31.5% whereas that of the incoherent matrix of the nodule-bed is 56.6%. A careful search has failed to reveal any borings in the nodules or any organisms encrusting them. Furthermore, the surfaces of the nodules are not phosphatised and are never stained green or brown. It is therefore an open question whether (1) the nodules are pebbles produced by the penecontemporaneous erosion of sandy limestone or (2) the matrix of the nodule-bed is incoherent because of partial decalcification, the nodules being undecalcified remnants of the original deposit. It is also possible that the nodules were produced by the spheroidal weathering of sandy limestone such as can be seen in some of the masses of Beer Stone in the old walls and buildings around Beer. No examples of the concentric flaking associated with such weathering has, however, been seen associated with the nodules in the Bovey Lane sand-pit.

There is no evidence of an erosion-plane above the nodular bed although the contact with the overlying glauconitic beds is sharp and well defined. Although it is difficult to prove that the nodules in the nodule-bed are water-worn pebbles it is possible that there was no important break between the deposition of non-glauconitic and glauconitic beds. The nodules of limestone could be pebbles derived from some adjacent source during the erosion-period suggested by the erosion-plane which separates Division A and Division B in the coastal sections. In other words, the beds in the Bovey Lane sand-pit appear to have accumulated in a depression or basin similar to that between the Beer Beach and Branscombe Mouth ridges. The small thickness of the Cenomanian Limestone in the poor exposures along Quarry Lane, between Hooken Cliff and Bovey Lane, suggests that the two depressions were not parts of one continuous belt.

The abundance of sand in Division C in the Bovey Lane sand-pit contrasts strongly with the base of the Middle Chalk exposed in the quarry on the south side of Quarry Lane. By analogy with Hooken Cliff the abundance of sand would appear to indicate the erosion of local sandy deposits during the early stages of the Turonian transgression. It is possible that the sand was derived from earlier Cenomanian deposits, or even from the top of the Upper Greensand, exposed in a ridge on the late Cenomanian and early Turonian sea-floor. There are no exposures in the vicinity of Bovey Lane which provide evidence of the location of such a ridge but it is doubtful whether it had any connection with that at Branscombe Mouth.

The relationship between the glauconitic deposit and the soft white chalk of the *Inoceramus labiatus* Zone which is exposed in the old overgrown quarries on the north side of Bovey Lane cannot be determined, but the irregular nature of the floor of Bovey Coppice, a wood *immediately* above the Bovey Lane sand-pit, suggests that chalk was once quarried there.

(c) The Branscombe District

Several inland exposures of the Cenomanian Limestone occur in the Branscombe district. They show that the lithology and thickness of the Formation are as variable as in the cliff-section, indeed all the exposures which have been examined in this area show different successions. The relationship between the various successions is, however, obscure because of the widely scattered nature of the exposures and the small size of most of them. The exposure in Stockham's Hill quarry (Fig. 5), about three-quarters of a mile ENE. of Branscombe Church, has already been described (p. 108). Another exposure of the Cenomanian Limestone occurs at the northern end of a small quarry on Margell's Hill (30/199890), about 700 yards north-east of Branscombe Church (Fig. 5). The exposure is very small and reddened by farm bonfires:

MIDDLE CHALK (? *I. labiatus* Zone). Hard white chalk containing glauconitic patches at or near the base, fossils rare apart from cidarid spines, much disturbed by roots from boundary-hedge. (2–3 feet.)
CENOMANIAN LIMESTONE.
Division B. (Absent.)
Division A 2. Hard shelly sandy limestone. (6 inches.)
Division A 1. Pebbly calcareous grit containing fragments of *Pecten* and *Neithea*. (8 inches.)
UPPER GREENSAND. Calcareous sandstone, erosion-surface at top. (3 feet seen.)

Poor incomplete exposures of the Cenomanian Limestone occur below the gated road along the western side of Woodhouse Hill. One of these, near the ruins of an old lime-kiln, about 250 yards south of the northern gate, shows a few inches of coarse calcareous grit from which a large *Ceriopora ramulosa* Michelin was collected. The exposure is at the top of a grassy bank so that it is not possible to determine the total thickness of the Cenomanian Limestone at this locality. The large quarries on Woodhouse Hill are now completely overgrown and any former exposures of the Cenomanian Limestone are now covered by scree and vegetation.

At the south-eastern end of Culverhole Hill there are two interesting quarry-sections which may be reached by ascending the steep Deepway Lane which runs past Branscombe School (Fig. 5). The first is at the top of the grassy slope a few yards past the gate at the end of Deepway Lane and overlooking the footpath leading to Hole Coppice (30/194889). This quarry has been excavated in cambered strata and includes three huge slipped blocks which have moved downhill into the valley of the Branscombe stream. The block forming the southern part of the working face exposes six feet of Upper Greensand and about two feet of very hard, compact white limestone containing scattered grains and pebbles of quartz but very little glauconite. Apart from a few shell-fragments no fossils have been seen in this limestone. It resembles most closely the thin development of the Cenomanian Limestone in the cliffs at Donkey Linhay and may therefore be a development of Division A 2 of the Cenomanian Limestone. It is not possible to determine whether Division B is present between this limestone and the Middle Chalk because of the dense vegetation and matted growths of ivy at the top of the pit.

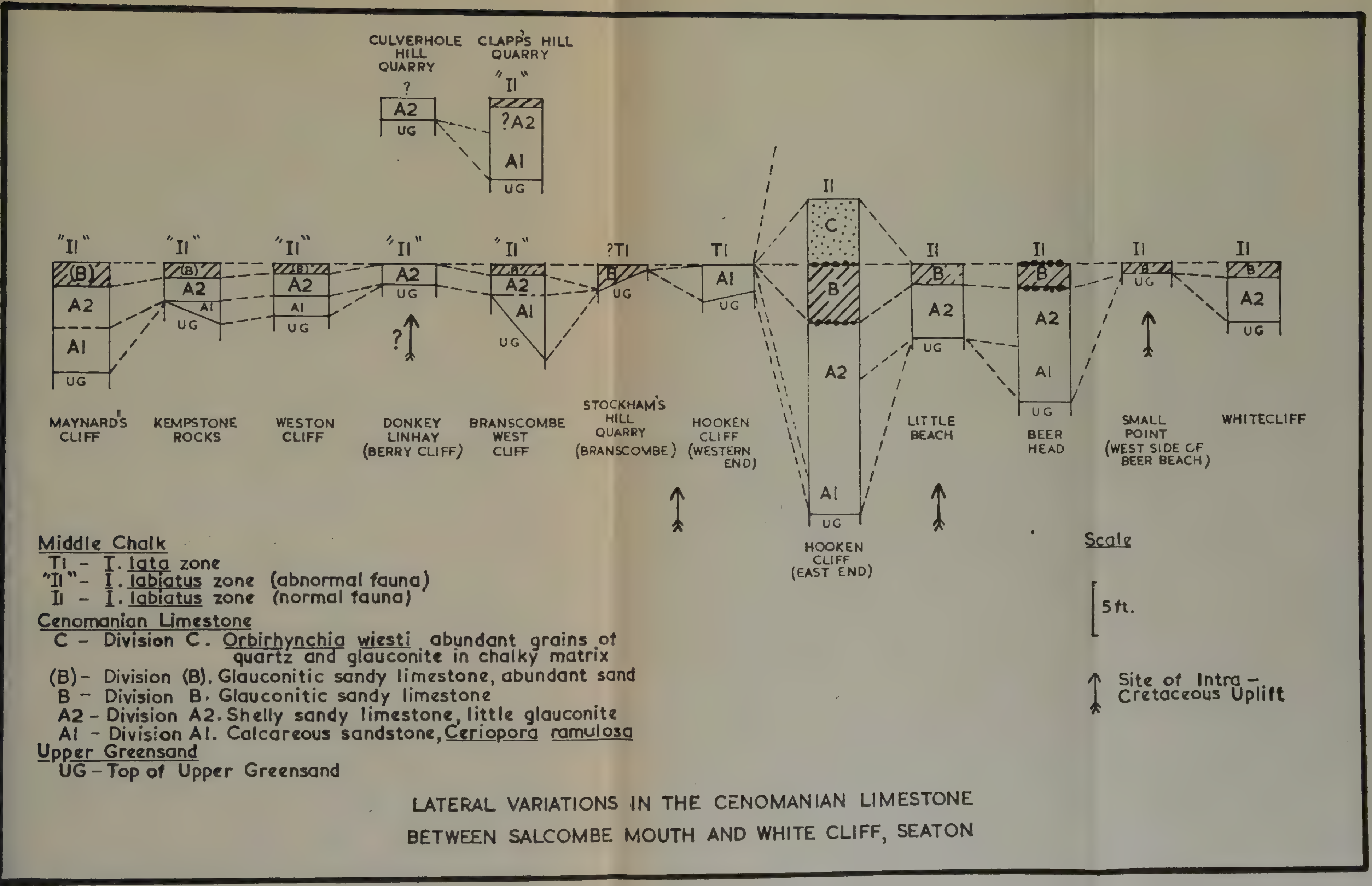
There are many overgrown chalk quarries at the summit of Culverhole Hill. Any former exposures of the Cenomanian Limestone in the floors of these quarries are now completely covered by scree and dense growths of brambles. A section of Cenomanian Limestone can be seen, however, in a badly overgrown quarry which could be called 'Clapp's Hill Quarry'. This quarry can be reached by ascending the grassy track leading southwards from the gate at the end of Deepway Lane (30/194888).

The face of this quarry is about twenty-five feet high but is hidden behind a dense growth of particularly high bramble bushes and is so steep that the beds in the upper part of the pit are inaccessible. Rough eye-measurements, however, confirm the thicknesses recorded by Jukes-Browne (1903), viz.:

		ft.	in.
TURONIAN	Rubbly soil and whitish nodular chalk	2	0
	Hard yellowish limestone, nodular at base	1	0
CENOMANIAN LIMESTONE	B. Hard compact whitish gritty limestone with green-coated nodules at top	0	9
	A. Rough gritty and pebbly rock, with casts of shells and sandstone pebbles, *Ceriocava* [*Ceriopora*] *ramulosa* near base	6	6
UPPER GREENSAND	Massive calcareous sandstone	15	0

Division A is not only very thick but is very coarse-grained, especially at the base. Jukes-Browne correlated the whole thickness with his Division A 1 as developed in Branscombe Cliff since there is no conspicuous boundary within the bed such as occurs between Divisions A 1 and A 2 in the adjacent cliffs. The present author is not in complete agreement with Jukes-Browne. The small exposures in the grassy slopes at the southern end of the quarry, and which form a continuation of the above section, suggest that the top of Division A in the quarry is a sandy limestone similar to Division A 2 on the coast. It is more likely therefore that there is a complete passage here between a particularly coarse-grained development of Division A 1 and more or less normal, but partially decalcified, Division A 2.

The section just described is only about 100 yards SSE. of the quarry opened in slipped blocks and in which the Cenomanian Limestone is thin and incomplete. It is evident that changes in the total thickness of the formation occur just as rapidly inland as they do in Branscombe West Cliff and Berry Cliff. The number of exposures in the Branscombe district


CULVERHOLE HILL QUARRY
?
A2
UG
CLAPP'S HILL QUARRY
"I1"
?A2
A1
UG
"I1"
(B)
A2
A1
UG
MAYNARD'S CLIFF
"I1"
(B)
A2
A1
UG
KEMPSTONE ROCKS
"I1"
(B)
A2
A1
UG
WESTON CLIFF
"I1"
A2
UG
?
DONKEY LINHAY (BERRY CLIFF)
"I1"
B
A2
A1
UG
BRANSCOMBE WEST CLIFF
?T1
B
UG
STOCKHAM'S HILL QUARRY (BRANSCOMBE)
T1
A1
UG
HOOKEN CLIFF (WESTERN END)
I1
C
B
A2
A1
UG
HOOKEN CLIFF (EAST END)
I1
B
A2
UG
LITTLE BEACH
I1
B
A2
A1
UG
BEER HEAD
I1
B
UG
SMALL POINT (WEST SIDE OF BEER BEACH)
I1
B
A2
UG
WHITECLIFF
Middle Chalk
T1 - I. lata zone
"I1" - I. labiatus zone (abnormal fauna)
I1 - I. labiatus zone (normal fauna)
Cenomanian Limestone
C - Division C. Orbirhynchia wiesti abundant grains of quartz and glauconite in chalky matrix
(B) - Division (B). Glauconitic sandy limestone, abundant sand
B - Division B. Glauconitic sandy limestone
A2 - Division A2. Shelly sandy limestone, little glauconite
A1 - Division A1. Calcareous sandstone, Ceriopora ramulosa
Upper Greensand
UG - Top of Upper Greensand
Scale
5 ft.
Site of Intra-Cretaceous Uplift

LATERAL VARIATIONS IN THE CENOMANIAN LIMESTONE BETWEEN SALCOMBE MOUTH AND WHITE CLIFF, SEATON

Fig. 8. N.B.—The NNE.–SSW. axis described by the author (Smith, 1957a, fig. 1) is crossed twice by these sections—at Little Beach and on the west side of Beer Beach

is far too small to show whether there is any connection between the thin incomplete developments of the Cenomanian Limestone at Donkey Linhay and that on the eastern flank of Culverhole Hill, or between the thick developments of this formation at The Cove and in Clapp's Hill quarry.

There are no clear inland exposures of the Cenomanian Limestone west of Branscombe. Several small rectangular blocks of white compact limestone lie in the Dunscombe Manor caravan-site and have presumably come from the old overgrown quarry east of Dunscombe Farm. The section formerly exposed in this quarry has been described by Jukes-Browne (1903, 142–3). His description shows that both Division A 1 and Division A 2 are present at this locality, the thickness of the former varying from fifteen to thirty inches. Division B appears to be exceptionally thin at this locality, being represented by only a layer of green-coated nodules. Within a few yards, however, the division thickens and assumes the lithology which is characteristic of its outcrops in the Dunscombe cliffs, viz. a whitish gritty limestone.

Small fragments and projecting scars of Cenomanian Limestone occur in the grassy slope above the quarry in the Upper Greensand at Lincombe, about half a mile south of Dunscombe Farm (Fig. 5). The western limit of the Cenomanian Limestone, which passes inland from Maynard's Cliff, must occur near the top of the eastern side of the valley south of Salcombe Regis. There are, however, no exposures in this region so that it is not possible to determine whether the exceptional thickness of the Cenomanian deposits at Maynard's Cliff is maintained inland.

6. SUMMARY

The principal lateral variations in the lithology and thickness of the Cenomanian deposits in the section described in the present paper and those previously described by the author (Smith, 1957a) are summarised in Fig. 8.

7. DISCUSSION OF THE FIELD EVIDENCE

(a) The Branscombe Mouth Ridge

(*i*) *Inadequacy of Previous Hypotheses.* Any explanation of the lateral variations in the thickness of the Cenomanian deposits between Hooken Cliff and Maynard's Cliff must demonstrate the nature and origin of the submarine ridge which undoubtedly existed in the Branscombe Mouth region during Cenomanian and early Turonian times. Jukes-Browne (1903, 139 and 440) assumed that the thick Cenomanian deposits at the eastern end of Hooken Cliff originally extended westwards to Branscombe Mouth, the thickness of these deposits possibly increasing in that direction so that they formed 'a kind of sand-bank'. He accounted for the absence of this thick sequence of Cenomanian deposits at the western end of Hooken Cliff today by suggesting that the whole of Division B and most of Division A were removed from this region by the action of marine currents during at least two intra-Cenomanian erosion-periods. Jukes-Browne (*op. cit.*, 440) also regarded the lower part of the Middle Chalk in Hooken Cliff as thinning out westwards against a submarine bank or ridge. His diagram (*op. cit.*, fig. 80, p. 441) suggests that he suspected that this ridge was of tectonic origin but he makes no reference to intra-Cretaceous earth-movements in his text. Jukes-Browne does not appear to have appreciated the full significance of tectonic uplift since he found it difficult to understand how the tract from which the Cenomanian deposits were almost entirely removed continued to form a submarine ridge during the accumulation of the lower part of the Middle Chalk (*op. cit.*, 440). He did not explain why relatively severe erosion should have occurred more than once in a similar location and did not account for the effects of current-scour being so much more severe immediately east of Branscombe Mouth than at the eastern end of Hooken Cliff, only a short distance away.

The present author does not consider that thick Cenomanian deposits accumulated in the Branscombe Mouth region. The thick Cenomanian deposits at the eastern end of Hooken Cliff, especially the greater part of Division B, contain only small amounts of sandy detritus. The submarine ridge in the Branscombe Mouth region could not therefore have been a sand-bank composed of Cenomanian sediments. At the time of their deposition these sediments were presumably unconsolidated calcareous muds containing scattered sand-grains; it is very unlikely that a submarine ridge could have been formed by the local deposition of an exceptional thickness of such sediments. Furthermore, although the high sand-content and derived fossils in Division C provide evidence of the erosion of earlier Cenomanian deposits, the small thickness and restricted distribution of Division C, and the small amount of sandy detritus in the Middle Chalk, show that no big thickness of sandy Cenomanian deposits could have been removed from the Branscombe Mouth region during intra-Cretaceous erosion. It is impossible to determine the original thickness of the Lower Cenomanian beds laid down at the western end of Hooken Cliff between the successive intra-Cenomanian erosion-periods since only remnants are still preserved. Such evidence as exists, particularly that provided by the section on Stockham's Hill, suggests that it is very unlikely that the total thickness of these Cenomanian beds ever exceeded that of the equivalent strata in the vicinity of the Beer Stone adit. The beds are much more likely to have thinned towards the Branscombe Mouth region in a similar manner to that which can be observed when they are traced towards the intra-Cenomanian submarine ridge at Beer (Smith, 1957a). The present author agrees with Jukes-Browne in ascribing the thinning-out of the lower part of the Middle Chalk in Hooken Cliff as almost entirely due to deposition against a submarine ridge. It is, however, impossible to determine whether Divisions A and B originally thinned in a similar manner, the observed thinning being almost entirely due to penecontemporaneous erosion.

Rowe (1903) did not include a description of the Cenomanian Limestone in his detailed account of the lateral variations in the Chalk of Devon. He gave a detailed description of the zonal overlap in the Middle Chalk at Hooken Cliff and referred to a glauconitic chalky bed containing a mingled Cenomanian and White Chalk fauna, presumably that termed 'Division C' in the present account. He concluded that 'both the factors of erosion and cutting off of beds by deposition against a Greensand bank must be employed to explain the whole chain of phenomena as they apply to the basal beds of the White Chalk' (*op. cit.*, 24).

Rowe did not attempt to account for the presence of this 'Greensand bank' but he seems to imply that a sand-bank produced by the deposition of an abnormal thickness of Upper Greensand existed in the Branscombe Mouth region during early Turonian times. The base of the Upper Greensand in Branscombe East Cliff and Hooken Cliff is concealed by the slipped blocks in the undercliff, but easterly projection of the base in Branscombe West Cliff suggests that the thickness of the Upper Greensand at the western end of Hooken Cliff is about 175 feet, a thickness which approximates to that at Whitecliff. A regional study of the top of the Upper Greensand shows that a bed of coarse greensand, about one foot thick, is persistent throughout the sections between Whitecliff and Dunscombe Cliff, west of Branscombe Mouth. Table I shows the vertical distance between the top of this bed of greensand and the base of the Cenomanian Limestone at six localities between Seaton and Branscombe West Cliff.

TABLE I

(1) Whitecliff (King's Hole)	11 feet
(2) Pound's Pool Beach	12 feet
(3) Little Beach (Beer Head)	10 feet
(4) Hooken Cliff (eastern end)	12 feet
(5) Hooken Cliff (western end, measured in slipped block, a few yards east of Mitchell's Rock)	13½ feet
(6) Branscombe West Cliff	11 feet

It is evident that the thickness of the Upper Greensand above the bed of coarse greensand is practically constant, despite the fact that the thickness of the overlying Cenomanian Limestone varies considerably. Furthermore, the bed of coarse greensand occurs at about the same distance below the base of the Cenomanian Limestone at the western end of Hooken Cliff, a locality which lay near the crest of the Branscombe Mouth Ridge, as at localities which were presumably low-lying areas of the sea-floor adjacent to the ridge. Although the pebbly sandstones immediately below the Cenomanian Limestone at Stockham's Hill are not developed in the more easterly sections, their thickness is much too small to account for a conspicuous feature on the Cretaceous sea-floor. Summing-up, therefore, it is clear that the Branscombe Mouth Ridge was not produced by the accumulation of a particularly great thickness of either Albian or Cenomanian sediments.

(*ii*) *The Effects of Differential Erosion.* The scour-channels in the surface of the Upper Greensand at Stockham's Hill show that minor irregularities on the Cenomanian sea-floor were produced by the erosive action of marine currents. The similarity in the lithology and thickness of the Upper Greensand between Whitecliff and Branscombe West Cliff shows, however, that a submarine ridge capable of causing the overlap of over seventy feet of Middle Chalk could not have been produced by differential erosion of the sea-floor during Cenomanian times. The thick Cenomanian deposits and the lower part of the Middle Chalk at the eastern end of Hooken Cliff were not deposited in a particularly deep scour-channel.

(*iii*) *Tectonic Uplift as the Dominant Factor.* Assuming therefore that the Branscombe Mouth Ridge was not due to either irregular deposition or irregular erosion of Albian and Cenomanian sediments the present author considers that it must have been of tectonic origin. The Albian and Cenomanian deposits immediately east of Branscombe Mouth must have been periodically uplifted by earth-movements to form a belt of particularly shallow water. Each upward movement would produce a submarine ridge over which the erosive action of marine currents would be more intense than in the adjacent areas of deeper water. In the uplifted region sedimentation would, for a time, be hindered or prevented. Such upward movements undoubtedly occurred during Lower Cenomanian times, but the greatest uplift must have taken place during Upper Cenomanian times just previous to the beginning of the accumulation of Division C in the deeper water east of the submarine ridge resulting from the uplift. The thickness of Middle Chalk overlapped in Hooken Cliff suggests that the Upper Greensand and Lower Cenomanian deposits in the Branscombe Mouth region were elevated by at least seventy-five feet during this late Cenomanian tectonic activity.

It should be noted that the postulation of a tectonic origin for the Branscombe Mouth Ridge avoids the difficulties associated with the views of Jukes-Browne and Rowe. For example, such an origin accounts for the estricted location of the area of periodic, relatively severe, penecontemoraneous erosion of Cenomanian sediments. It also explains why the verlap of a considerable thickness of Middle Chalk should be associated ith only minor variations in the thickness of the pre-Turonian sediments.

(*iv*) *The Nature of the Tectonic Uplift in the Branscombe Mouth Region.* he determination of the exact nature of the tectonic uplift which produced e Branscombe Mouth Ridge is hindered by the destruction of critical ections by post-Cretaceous erosion. At Branscombe West Cliff the retaceous rocks have a gentle easterly dip and the top surface of the pper Greensand crops out at about 350 feet above sea-level. The valley f the Branscombe Stream breaks the continuity of the cliff-section of the pper Greensand for nearly three-quarters of a mile and the top surface e-enters the cliff-section at Branscombe East Cliff. Here it occurs at a imilar altitude to that at Branscombe West Cliff, i.e. higher than would be xpected if the easterly dip of the Cretaceous rocks at the latter locality as maintained across Branscombe Mouth.

Minor displacements due to cambering of the top beds of the Upper reensand are common around Branscombe but there is no evidence of a ajor fault in the Cretaceous rocks which could account for the raising of e Upper Greensand in Branscombe East Cliff by at least seventy-five feet. he periodic uplift of the Albian and Cenomanian rocks to the east of ranscombe Mouth does not, therefore, appear to have been the result of enecontemporaneous faulting but is much more likely to have been the esult of movements along an anticlinal axis trending in a general northerly irection from the coast near Branscombe Mouth.

The Middle Chalk at the extreme eastern end of Branscombe West Cliff ncludes twenty feet of nodular chalk which, although possessing an bnormal fauna (Rowe, 1903, 25), can be referred to the *Inoceramus biatus* Zone. The reappearance of this zone just under a mile west of the oint at which it is overlapped in Hooken Cliff indicates that the lower part f the Middle Chalk at Branscombe West Cliff must have accumulated on e western side of the Branscombe Mouth Ridge. None of the Cretaceous ocks in Branscombe West Cliff shows a westerly dip, however, so that it is ossible that the whole of the western limb of the anticline responsible for e ridge, together with its easterly-overlapping series of Middle Chalk eds, have been removed by the excavation of the valley of the Branscombe tream. Alternatively the anticline may have been asymmetrical so that the bmarine ridge sloped westwards at an angle less than the present-day asterly tilt of the Cretaceous strata as a whole. In this respect it should be oted that the eastern flank of the Branscombe Mouth Ridge cannot be escribed as steep, a gradient of about 1 in 23 being sufficient to account r the overlap of Middle Chalk beds observed in Hooken Cliff. It is ossible that during early Turonian times the sea-floor west of Branscombe

Mouth sloped gently, without interruption, at least as far west as Berry Cliff. Here, the *Inoceramus labiatus* Zone is said by Rowe (1903, 37) to include up to sixty feet of flinty chalk in addition to the more norma flintless chalk, the total thickness of the zone reaching eighty feet, a thickness far exceeding that at any other locality on the coast of south-east Devon. If this thickness of the *Inoceramus labiatus* Zone can be accepted it would be in keeping with deposition on the western side of an asymmetrica anticlinal fold with its crest in the Branscombe Mouth region.

Little can be said with regard to the nature of the western side of the Branscombe Mouth Ridge during the accumulation of the Cenomanian Limestone. The prevalence of scour-channels at the top of the Upper Greensand and the particularly uneven erosion-surfaces of the Cenomanian beds in Branscombe West Cliff and Berry Cliff could be due to the increased erosive action of marine currents being forced to flow over a sea-floor rising gently eastwards towards the crest of a submarine ridge and facing the direction from which the currents entered the Branscombe district. It is therefore significant that the scour-channels at the top of the Upper Greensand extend in a general E.–W. direction, approximately at right-angles to the presumed strike of the axis of uplift in the Branscombe Mouth region.

(*v*) *The Northern Extent of the Branscombe Mouth Ridge*. The scarcity o exposures in the Branscombe district makes it difficult to determine the northern extent of the Branscombe Mouth Ridge. The presence of the *Inoceramus labiatus* Zone in the old quarries north and north-east o Branscombe shows that the zonal overlap towards the crest of the ridge extended for only a short distance north of Stockham's Hill, presumably because of northerly plunge of the anticlinal uplift responsible for the Branscombe Mouth Ridge. The relatively great thickness of Cenomanian deposits and the absence of evidence of severe intra-Cenomanian erosion in the Bovey Lane sand-pit also indicate a general northerly plunge of the anticlinal uplift. A comparable example of this phenomenon occurs between Beer Head and Hooken Cliff where the appearance of thick Cenomanian deposits has been ascribed by the author to southerly plunge of an intra-Cretaceous axis of uplift (Smith, 1957a). The sands in the Bovey Lane section could not have been deposited as a sand-bank rising above the general level of the Cenomanian sea-floor since there is no evidence of overlap of younger beds towards them. The sands must have been deposited in a low-lying area of the Cenomanian sea-floor and were not washed away, as was the case elsewhere, because they lay below the general level of erosion.

It is highly probable that the periodic uplifts along the line of the Branscombe Mouth Ridge occurred as the result of the development of an elongated pericline, possibly of asymmetric type, and with an axis elongated in a general N.–S. direction. Complete proof of this hypothesis is rendered impossible by the destruction of the southern part of the Branscombe uplift by coastal erosion.

(*vi*) *The Branscombe Mouth Ridge as a Factor Responsible for the Anomalous Fauna of the* Inoceramus labiatus *Zone West of Branscombe Mouth.* Rowe (1903, 25) described the faunal anomalies of the lower part of the Middle Chalk in the cliffs west of Branscombe Mouth (Fig. 8). He noted the absence of *Inoceramus mytiloides* (=*labiatus*), *Conulus castanea*, *Cardiaster cretaceus* and the scarcity of '*Rhynchonella*' *cuvieri*, *Discoidea dixoni*, *Hemiaster minimus* and *Cardiaster pygmaeus*, all of which are characteristic of the *Inoceramus labiatus* Zone in the section east of Branscombe Mouth. Rowe also commented upon the exceptional abundance of spines and plates of cidarids and the ossicles of pentacrinids and asteroids in the sections west of Branscombe Mouth. Wright (in discussion, Smith, 1957a) suggested that these faunal contrasts were due to a submarine ridge to the west of Hooken Cliff having acted as a barrier, presumably implying that the ridge hindered the migration of marine organisms.

The present author considers that the Branscombe Mouth Ridge did not act as a barrier preventing the migration of organisms such as echinoids. The ridge was undoubtedly the site of particularly shallow water during early Turonian times but since it appears to have been of no great lateral extent migrating organisms could easily have avoided the shallows by keeping to the deeper water surrounding the ridge. It is much more likely that the faunal contrast is due to palaeoecological differences on the early Turonian sea-floor. The depth of water, the strength and direction of currents and variations in the distribution of small organisms which formed the food-supply of the larger organisms would all influence the distribution of life on the Turonian sea-floor. An asymmetrical submarine ridge and the adjacent areas of deeper water would obviously provide a wide range of environments lying at varying depths. Furthermore, the sea-floor on the western side of the Branscombe Mouth Ridge would be more exposed to marine currents flowing towards the east than the more-sheltered eastern side in the area of Hooken Cliff. The opposite sides of the ridge could therefore provide the home for contrasting faunal assemblages although some organisms, not requiring particularly critical conditions, would be more widespread. This conclusion is supported by the fact that the base of the Middle Chalk in some of the large fallen blocks at the western end of Little Beach contains abundant cidarid spines and plates. Evidently the conditions suitable for the development of large colonies of cidarids did exist on the eastern side of the Branscombe Mouth Ridge for a short period but, judging by the smaller frequency of cidarid spines in the remainder of the zone, these conditions did not persist for as long as they did on the western side of the ridge.

(b) Sedimentation Problems Associated with the Cenomanian Deposits at the Eastern End of Hooken Cliff

It is difficult to account for the uneven distribution of sand in the upper part of Division A and Division B, particularly the sandy lenticles in the upper part of Division B at Hooken Cliff and in the Bovey Lane sand-pit (Fig. 4a). It is unlikely that the latter were produced by sand penetrating into a complex series of borings or burrows (Hancock, in discussion, Smith, 1957a) since it is only when the lenticles increase in frequency that any vertical connection between them becomes apparent. The limestone between the sandy lenticles does not appear to have been disturbed so that the penetration of sand between displaced blocks is not a likely explanation. The author considers that the lenticles were produced by the periodic irregular winnowing of patchily consolidated calcareous mud by marine currents which would be eroding the adjacent ridges more vigorously and could have carried in sand derived from these ridges. In other words, the lenticles can be interpreted as miniature sand-filled scour-channels, their lateral distribution at successive levels being controlled by lateral variation in the cohesion of the calcareous mud swept by the marine currents. These currents were evidently not persistent but occurred at more frequent intervals as the deposition of calcareous mud progressed. When one period of current-activity was followed rapidly by another it would be possible for scour-channels to cut down into earlier ones, thus giving the vertical coalescence of lenticles found in the upper part of the sequence. Eventually the strength of the currents became sufficiently intense to displace the more consolidated patches of calcareous mud to a minor extent, the displaced patches being rounded and enveloped in calcareous mud enriched with sand brought in by currents entering the Hooken area. In this way one can account for the nodular deposit immediately beneath the pebble-bed at the top of Division B. A similar mechanism could account for the nodular character of the upper part of Division A in which the rounded masses of limestone are distinctly less sandy than the sediment enclosing them, and are therefore unlikely to be concretionary.

The pebble-beds at the top of Division A and Division B indicate that in both cases the currents eventually became sufficiently intense to transport small masses of consolidated limestone over greater distances than hitherto, for example, from the crests of the adjacent submarine ridges. The pebbles, once they had come to rest, appear to have lain on the sea-floor for a considerable time before they were buried beneath newer sediments, during which time they were superficially phosphatised. That this occurred *in situ* is suggested by the fact that in places the pebble-bed has a phosphatic veneer which coats both pebbles and the matrix between them. The pebble-beds are clearly not basal-conglomerates but may be regarded as 'terminal-conglomerates' since each represents a period of inhibited sedimentation rounding off a cycle of sedimentation. It is difficult to account for the cessation of deposition suggested by the phosphatised pebble-beds, particularly over the long period indicated by the major stratigraphical break between Division B and Division C. Any Upper Cenomanian sediments deposited during this time-interval must have been completely swept away from the area under consideration before the accumulation of the basal part of Division C and the Middle Chalk. Possibly this part of the Upper Cenomanian sea-floor was more or less persistently swept by currents, the velocity of which, although not sufficient to disturb the pebbles or to transport sand-grains, was sufficient to prevent the accumulaion of fine-grained chalky sediment. This suggestion is supported by the fact that in some instances the pebbles are truncated by the upper surface of Division B, indicating some erosion of the pebble-bed after the conolidation of its matrix. The erosion-surface in such cases is, however, uperficially phosphatised showing that the erosion occurred during an nterval in a lengthy period of phosphatisation.

The long period of inhibited deposition suggested by the almost complete absence of Upper Cenomanian sediments in the sequence may also have been due to the temporary cessation of tectonic activity. After uplift of the ubmarine ridges during the closing stages of the accumulation of Division B, the Cenomanian sea-floor was levelled by erosion of the ridges and the ccumulation of sediments, including spreads of pebbles, in the intervening depressions. By this time there would be no upstanding areas which could e preferentially attacked by marine currents and no sheltered areas in which unconsolidated sediments could accumulate. Marine currents could herefore prevent the accumulation of sediment over the whole of the Beer nd Branscombe district. Towards the end of Cenomanian times, however, he sea-floor was again rendered uneven by renewed uplift of the Bransombe Mouth Ridge. Unconsolidated sandy sediment (Division C) could hen accumulate in a depression on the sheltered side of the ridge although marked slackening of current-activity must have been an important actor responsible for the accumulation of the relatively thick deposits of Middle Chalk which eventually levelled the Turonian sea-floor by burial of he Branscombe Mouth Ridge.

c) Factors Responsible for the Lateral Variations in the Thickness of the Cenomanian Beds between Berry Cliff and Maynard's Cliff

The poor and widely scattered nature of the exposures between Berry Cliff and Maynard's Cliff and those north of the cliff-line make it difficult o determine whether any submarine ridges of tectonic origin developed est of the Branscombe Mouth Ridge. The lateral variations in the total ickness of the Cenomanian Limestone in this region appear to be mainly due to the presence of undulating erosion-surfaces, especially that forming the top surface of the Upper Greensand.

It is, however, possible that a minor area of periodic tectonic uplift was sited in the Donkey Linhay region since,

(i) The Cenomanian Limestone is here thinner and less complete than in the cliffs to the east and west,

(ii) The blocky Upper Greensand over The Cove, with Cenomanian sediments between the blocks, is very much like that near the crest of the submarine ridge on the west side of Beer Beach (Smith, 1957a),

(iii) The lithology of Division B in the sections west of Donkey Linhay differs from that in the Branscombe cliffs, to the east.

None of these features considered by itself provides conclusive evidence of tectonic uplift, but collectively they are of greater significance. Nevertheless, with the evidence at present available, the existence of a third area of intra-Cenomanian earth-movements in the coastal region west of Seaton cannot be regarded as firmly established.

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DISCUSSION

DR. J. M. HANCOCK congratulated the author on this further valuable contribution of his detailed researches on the Cretaceous of Devon. The ammonites from this region confirmed both the appreciable time-gap between Beds B and C and the great condensation in the fauna at the base of Bed C. This was probably equivalent to two, three or more subzones on the scale of the zoning of the Gault. Comparison of the ammonites with those of the type Cenomanian, showed that Bed C not only contained ammonites of the Marne à *Ostrea biauriculata* (e.g. *Eucalycoceras rowei* (Spath)), the highest formation of the type Cenomanian, but also forms such as *Euomphaloceras* and '*Utaturiceras*' which are probably still younger.

The speaker was most interested in the irregular distribution of the glauconite in the calcareous cement in Bed B in the section below the Pinnacles. The absence of vertical connections between the patches certainly disproved their representing organic burrows. Had the author considered the possibility that the irregular distribution in the scour-channels might be connected with differing cohesions resulting from the different thixotropies of the glauconite sand and the calcareous mud?

Would it be possible to use a hand-powered drilling rig to check the thickness of the Wilmington sand facies in the Bovey Lane pit?

DR. D. V. AGER commented on the interesting faunal differences in the Lower Turonian on either side of the Branscombe Ridge. He agreed that this was almost certainly an ecological matter and asked if there was any evidence of differences in the solidity of the sea-floor at this time. The dominant fauna on the west side—of 'regular' echinoids, crinoids and asteroids—suggested a firm substratum (and possibly a more protected habitat), whereas the Inocerami and 'irregular' echinoids east of the ridge suggested an unconsolidated floor in which it was possible to burrow. The presence of abundant rhynchonellids in the latter fauna might appear to be a problem, since it is usually assumed that pedunculate brachiopods need to attach themselves to a firm surface, but such occurrences are well known in many parts of the stratigraphical column and the speaker thought that other explanations were possible.

THE AUTHOR, in reply to Dr. Hancock, stated that he considered that most, if not all, of the ammonites in Division C were derived or *remanié* forms and great care is necessary in using them to give a precise age to the deposit. Nevertheless, it is clear from Dr. Hancock's careful study of the fauna that the bed must be, at the earliest, very late Cenomanian in age.

In reply to Dr. Hancock's second point the Author stated that he thought that the thixotropy of chalk-ooze could have been a factor in the formation of nodular deposits such as occur in the upper part of Divisions A and B. In heterogeneous mixtures produced by the disturbance of chalk-ooze by sand-laden currents, patches containing little or no sand would tend to 'set' earlier than the sandier patches by virtue of the higher thixotropy of the chalk-ooze. An assessment of the relative importance of such thixotropic transformations and those diagenetic changes loosely termed 'consolidation' is, however, very difficult. The possibility of using a hand-powered drilling rig, both in the Bovey Lane sand-pit and in the White Hart sand-pit at Wilmington, will be explored. In both cases it is highly desirable that the precise total thickness of the Cenomanian sands and the nature of the junction with the Upper Greensand be determined.

In reply to Dr. Ager the Author stated that he was very interested in a palaeoecological explanation of the lateral variations in the fauna of the Lower Turonian rocks in the Branscombe Mouth region. In early Turonian times the influence of easterly-flowing currents would be more pronounced on the western side of the Branscombe Mouth Ridge than on the more sheltered eastern side. The periodic formation of a firm sea-floor as a result of the sweeping away of unconsolidated chalk-ooze is therefore more likely to have occurred on the western side than on the eastern.

Field Meeting in the Gault at Small Dole, near Henfield, Sussex

Sunday, 14 August 1960

Report by the Director: R. A. MILBOURNE

Received 5 September 1960

THE PRINCIPAL object of this meeting was to examine and collect fossils from the Gault (Middle and Upper Albian) exposed in the British Portland Cement Company's clayfield at Horton, one-quarter of a mile south of Small Dole (51/–). A short visit was first made, however, to the Henfield Brick Company's working in Weald Clay, one mile south of Henfield (51/–). Both exposures are situated on the west side of the A.2037, Henfield to Upper Beeding road.

The Henfield Brick Co's. pit exposes some thirty feet of ferruginous clays, belonging to the Upper Weald Clay series, with, near the top, a thin seam of limestone crowded with shells of the freshwater gastropod *Viviparus* (*Paludina*) *fluviorum* J. Sowerby. This *Paludina* limestone, which is also known as Sussex Marble, yielded, as did the surrounding clay, bones and scales of the Wealden fish *Lepidosteus*. Large pieces of fossil wood in which the original cellulose has been mainly replaced by iron sulphide (pyrites) occurred on the weathered clay slopes, and employees of the company reported the finding there of large bones (presumably reptilian) during the summer of 1959.

On arriving at the Gault exposure at Small Dole, the party took advantage of a temporary halt in the proceedings, made necessary by a heavy shower of rain, to consume lunch, after which examination of the extensive workings commenced. This clayfield is being developed rapidly and is, without doubt, already the largest man-made Gault exposure in England.

The Gault, which is used together with Chalk in the manufacture of cement, is removed by multi-bucket excavator, working at the pit-side, and by bulldozers working across a large, flat arena. Conveyor belts transport the clay to a gigantic mixing and grinding tank from which it is pumped, as a high viscosity suspension in water, through twin pipelines to the company's Shoreham works at Beeding, two and a half miles to the south.

The pit exhibits some ninety feet of highly fossiliferous clays of which fifty-five feet are referable to the Lower Gault and thirty-five feet to the Upper Gault. Ammonites collected over a period of several years show that the lowest part of the section is in the upper *intermedius* Subzone of the *dentatus* Zone, and the highest part in the *orbignyi* Subzone of the *inflatum* Zone. Ammonite horizons present, together with their thicknesses, are as follows:

Zones	Subzones	ft.	in.
Mortoniceras inflatum	*Hysteroceras orbignyi*	35	0
Euhoplites lautus	*Dipoloceras cristatum*	6	0
	? *Anahoplites daviesi* and/or *Euhoplites lautus* and *E. nitidus*	6	0
	New Subzone	18	0
Hoplites dentatus	*Dimorphoplites niobe*, including the horizon of *Mojsisovicsia subdelaruei*	9	3
	Anahoplites intermedius (pars)	17	0
	Total	91	3

The Gault at Small Dole shows several marked differences in the ammonite succession when compared with the Folkestone Gault, but the succession is essentially similar to that found in the northern part of the Weald. The expansion of the lower part of the *lautus* Zone is marked, and is even greater than that noted at Greatness Lane, Sevenoaks, Kent (Milbourne, 1955).

The new Subzone mentioned in the foregoing table is the expanded equivalent of the upper layer of phosphatic nodules in Bed IV and a few inches of immediately overlying clay in Bed V at Folkestone. The occurrence of this new Subzone in the Sevenoaks area, for example, has already been noted (Milbourne, 1955, 241), and the commonest and most characteristic ammonites occurring within it are *Dimorphoplites doris* Spath and the late mutation of *Euhoplites loricatus* figured by Spath (1928). At Small Dole the latest part of this new Subzone is further characterised by the occurrence, in large numbers of well-preserved individuals, of a new species of *Euhoplites*, intermediate in all characters between the late mutant of *E. loricatus*, mentioned above, and *E. lautus* of the next (higher) Subzone.

Mojsisovicsia subdelaruei and allies occur profusely in the upper part of the *niobe* Subzone at Small Dole, forming a distinct horizon, as at other localities in the Weald.

It is not yet possible to demonstrate the occurrence of either the *lautus*–*nitidus* or *daviesi* Subzones at Small Dole since the six feet of clay occurring between the new Subzone and the *cristatum* Subzone have, so far, yielded only *Anahoplites planus* (Mantell) and *Dimorphoplites chloris* Spath, both of which are long-range species.

The *cristatum* Subzone, at the junction of the Lower and Upper Gault, is represented by six feet of clay, very dark and shaly at the top, with large phosphatic nodules scattered throughout. The concentration of nodules and fossils in this clay is not as great as in the well-known, equivalent nodule bed at Folkestone (lower Bed VIII) and at Greatness Lane (bed 18), although it is probable that condensation of this clay would result in a nodule bed of similar aspect.

Ammonites characteristic of the *orbignyi* Subzone, i.e. Bed IX of the Folkestone succession, are not of frequent occurrence in the lower part of the Upper Gault at Small Dole; *Hysteroceras orbignyi* and allies occurring plentifully in the topmost part of the section only. This feature, combined with the absence of any species characteristic of the overlying *varicosum* Subzone, shows that all thirty-five feet of Upper Gault belong to the *orbignyi* Subzone. This Subzone is a little over nine feet thick at Folkestone where, as at all other localities, the lower part is easily recognised by the numerous examples of *Euhoplites inornatum* Spath which it contains.

Of some interest is the frequent occurrence of the uncoiled ammonoid *Psilohamites bouchardianus* (d'Orbigny) in the *orbignyi* Subzone at Small Dole. Dr. Spath (1941, 656) records this rather rare form only from the *varicosum* Subzone, but in the author's experience the species is confined in occurrence to the *orbignyi* Subzone at all localities.

In physical characters the Gault at Small Dole is unlike that at Folkestone (Jukes-Browne, 1900, 71–81) in the highly ferruginous nature of the clays, with three major and many lesser seams of clay–ironstone lenticles, and in the dearth of concentrated layers of phosphatic nodules with related increase in the thickness of all horizons.

Many fine fossils were obtained by the party during the course of the afternoon and particular interest centred around a patch of clay with associated ribs, vertebrae and other bones of a large marine reptile. (Dr. W. Swinton of the British Museum (Nat. Hist.) has identified these bones as crocodile type.) Smaller fossils of interest included several fish otoliths and teeth, also many well-preserved lamellibranchs, gastropods, crustaceans and ammonites; the iridescent, pyritised examples of the latter prompting Dr. Hancock to remark that collecting conditions and mode of preservation of specimens were undoubtedly closely comparable with those obtained on the foreshore at Folkestone during the last century.

A sufficiently large number of car owners were present to provide transport for all to and from Henfield Railway Station. The advertised double four mile walk was thus happily avoided and a considerable increase in time available for study of the sections was obtained.

After a vote of thanks to the Director from Dr. Hancock, the party, which numbered twenty-two, dispersed in quest of tea.

REFERENCES

JUKES-BROWNE, A. J. 1900. The Cretaceous Rocks of Britain. 1, Gault and Upper Greensand of England. *Mem. geol. Surv. U.K.*

MILBOURNE, R. A. 1955. The Gault at Greatness Lane, Sevenoaks, Kent. *Proc. Geol. Ass., Lond.*, **66**, 235.

SPATH, L. F. 1923–43. A Monograph of the Ammonoidea of the Gault. *Palaeontogr. Soc.* (*Monogr.*).

R. A. Milbourne
77 Castle Drive
Kensing, Kent

The Annual Report of the Council of the Geologists' Association for the year 1960

MEMBERSHIP

HE numerical strength of the Association on 31 December 1960, was as ollows:

Honorary Members	22	(21)
Ordinary Members:		
Life Members (compounded) ...	189	(173)
Annual Subscribers	1732	(1863)
Institutional Members	60	(55)
	2003	(2112)

(Corresponding figures for 31 December 1959 in brackets)

During the year ninety-six new members were elected (including four nstitutional members) and the Association lost 206 members through leath, resignation and removal under Rule XI for failure to pay subscripions. One institutional member was reinstated.

The list of deceased members is as follows:

F. G. F. Barrington, W. D. Bell, Miss D. A. Bigby, P. G. H. Boswell, . W. S. Fawcett, J. H. Goodchild, F. Jones, W. A. Lambert, Mrs. L. V. Long, E. Montag and J. B. Simpson.

HOUSE LIST

he following retirements are announced:

Mr. D. Curry, as Senior Vice-President, Dr. J. M. Hancock, as Field Meetings Secretary, at his own request, Mrs. Carreck and Messrs. Elliott nd M. K. Wells, as ordinary members of Council.

Thanks are due to these members for valuable services rendered to the Association.

FINANCE

The year 1960 saw a considerable improvement in the Association's inancial position, as the result of the increase in the subscription rate on January 1960. Subscription income increased to £3097, but arrears outtanding at 31 December were considerably increased at nearly £600. Income from Advertisements', which in previous years had shown the ctual receipts from advertisers, now shows the net income after deducting he cost of printing the advertisements. The cost of printing the PROCEEDNGS is the highest ever at £2380, due to the increased size of the Volume (the largest since 1939) and to increased printing rates following the rise in printing wages in 1959. A substantial reduction in General Expenses has been secured.

Investment Income increased by about £100 mainly owing to the increased income of the Curry Fund. To help meet the increased liabilities in respect of Life Members resulting from the increased subscription rate, the income of the Compounding Fund, together with an additional sum of £100 from General Funds, has been added to the capital of the Fund. The balance of the surplus of General Funds on the year's working has been transferred to the General Reserve Fund.

Included in the Bequest Fund is a sum of £1840 16*s.* 3*d.* received by the Association as a residuary legatee of the late Mr. H. E. Busbridge, a Member who died in 1954. On the advice of the Trustees, most of this money has been invested in 5% Conversion Stock 1971. The Association also received a bequest of £200 under the will of the late Mrs. Llewellyn Treacher to be used in making grants for Pleistocene Research. This money has been placed in a separate Trust Fund. A first grant of £20 has been made from it.

Six Guides were published during the year; expenditure of £843 0*s.* 4*d.* recorded in the Centenary Guides Fund Account covers the cost of production of these and the residue of the cost of a Guide published in 1959. Production costs have sharply increased following the revision of the very favourable printing rates in force during the previous two years. Income from sales of Guides continued buoyant; initial sales of new Guides were good, and sales of the older Guides continued on the whole at a steady rate.

TRUSTEES

The Trustees of the Association were:

Managing: Dr. W. F. Fleet, M.SC., A.R.I.C., A.C.P., F.G.S., Mr. Percy Evans, M.A., F.G.S., and Dr. W. A. Macfadyen, M.C., M.A.

Custodian: The Royal Bank of Scotland, Western Branch.

ROYAL SOCIETY TERCENTENARY 1960

An Address of congratulation was presented to the President of The Royal Society by the Association as follows:

'On behalf of the Geologists' Association we send our congratulations on the occasion of your Tercentenary. During the three hundred years which have shown the greatest advances in knowledge you have been active in fostering science in its manifold branches and we wish you continued prosperity in the years ahead.'

At the formal opening ceremony of the Tercentenary Celebrations in the Albert Hall, the Association was represented by the President and the General Secretary.

BALANCE SHEET AT 31 DECEMBER 1960

Statement I

1959		CAPITAL		REVENUE	TOTAL
		Invested	*Uninvested*		
£		£ s. d.	£ s. d.	£ s. d.	£ s. d.
	FUNDS FOR SPECIAL PURPOSES:				
	Bequest Fund	4421 14 0	268 3 5		4689 17 5
	Curry Fund	2500 0 0			2500 0 0
	Compounding Fund	2507 9 7	211 0 8		2718 10 3
	TRUST FUNDS:				
	Foulerton Award Fund	294 11 0		17 0 6	311 11 6
	Henry Stopes Memorial Fund ...	260 12 9		34 6 8	294 19 5
	G. W. Young Geological Investigation Fund			42 10 11	42 10 11
	Treacher Fund for Pleistocene Research		180 0 0		180 0 0
8811		9984 7 4	659 4 1	93 18 1	10,737 9 6
	GENERAL FUNDS:				
183	Subscriptions paid in advance ...				165 10 0
211	Sundry Creditors				209 2 2
	Estimated Accrued Expenditure:				
400	Printing and Postage of PROCEEDINGS, Vol. 71, Part 4				570 0 0
35	Printing Monthly Circular ...				35 0 0
13	Reserve for Index of PROCEEDINGS				12 16 4
—	Advertising Income, held in Suspense				11 12 3
—	General Reserve Fund				558 6 6
					£12,299 16 9

1959			
£		£ s. d.	£ s. d.
8284	Investments		9984 7 4
188	Income Tax Recoverable		276 18 4
	Cash at Bank:		
	Current Account	407 17 0	
	Deposit Account	1300 0 0	
1121			1707 17 0
31	Petty Cash in hands of officers		18 16 5
55	Proceeds of sale of Guides in hands of Editor		40 7 0
100	Estimated Value of Arrears of Subscriptions (Book Value £583 10*s.*)		200 0 0
—	Cheque in hands of Treasurer		71 10 8
			£12,299 16 9

INCOME AND EXPENDITURE ACCOUNT FOR THE YEAR ENDED 31 DECEMBER 1960

Statement 2

EXPENDITURE

1959 £			£ s. d.	£ s. d.
54	DEFICIT AT 31 DECEMBER 1959 BROUGHT FORWARD			80 16 8
1582	PRINTING 'PROCEEDINGS' OF THE ASSOCIATION: (including £500 for estimated accrued expenditure for Vol. 71, Part 4)			2379 16 0
	GENERAL EXPENSES:			
39	Meetings (including teas)		71 6 10	
405	Printing Monthly Circulars (including estimated accrued expenditure of £35) ...		312 19 10	
	Printing 'Rules and List of Members'	351 0 1		
	Less charged in 1959 account	200 0 0		
200			151 0 1	
183	Printing and Stationery		198 19 1	
	Postages:			
299	PROCEEDINGS	283 9 7		
172	Circulars	114 7 6		
49	Miscellaneous	61 7 11		
			459 5 0	
59	Typing and Duplicating		33 9 11	
20	Purchase of past issues of PROCEEDINGS ...		4 1 0	
20	Incidental Expenses		44 16 4	
				1275 18 1
Deficit 81	Surplus at 31 December 1960:			
	Transfer to Compounding Fund		100 0 0	
	Balance to General Reserve Fund ...		558 6 6	
				658 6 6
				£4394 17 3

INCOME

1959 £		£ s. d.
1837	ANNUAL SUBSCRIPTIONS (Including £200 being estimated value of book arrears at 31 December 1960)	3097 8 5
90	ADMISSION FEES	65 10 0
	NET INCOME FROM ADVERTISEMENTS	20 11 2
342	SALES OF PUBLICATIONS (other than Guides)	465 16 9
—	INTEREST ON BANK DEPOSIT	15 9 11
86	INCOME FROM ILLUSTRATIONS FUND	20 10 6
38	INCOME FROM BEQUEST FUND	147 0 6
82	INCOME FROM COMPOUNDING FUND	— — —
475	INCOME FROM CURRY FUND	562 10 0
		£4394 17 3

CENTENARY GUIDES FUND

Statement 3

Income and Expenditure Account

	£	s.	d.		£	s.	d.
Printing six Guides and part cost of printing another	843	0	4	Income from Sales	513	19	11
				Balance to Working Capital Account	329	0	5
	£843	0	4		£843	0	4

Working Capital Account

	£	s.	d.		£	s.	d.
Net Expenditure to 31 December 1959	1024	6	0	Initial Capital	529	2	9
Net Expenditure in 1960	329	0	5	Advances from Bequest Fund	824	3	8
	£1353	6	5		£1353	6	5

BEQUEST FUND

Statement 4

	Investments £	s.	d.	Uninvested Capital £	s.	d.	Revenue £	s.	d.
New Investment				1600	0	0			
Advance of Capital to Centenary Guides Fund				329	0	5			
Transfer to Income and Expenditure Account							147	0	6
Balance at 31 December 1960	4421	14	0	268	3	5			
	£4421	14	0	£2197	3	10	£147	0	6

	Investments £	s.	d.	Uninvested Capital £	s.	d.	Revenue £	s.	d.
Balances at 1 January 1960	2821	14	0	346	7	7			
Bequests				1850	16	3			
New Investment	1600	0	0						
Investment Income							147	0	6
	£4421	14	0	£2197	3	10	£147	0	6

FUNDS FOR SPECIAL PURPOSES (*Continued*)

CURRY FUND

Statement 5

	Invested Capital £ s. d.	Revenue £ s. d.		Invested Capital £ s. d.	Revenue £ s. d.
Transfer to Income and Expenditure Account ...		562 10 0	Balance at 1 January 1960	2500 0 0	
Balance at 31 December 1960	2500 0 0		Investment Income		562 10 0
	£2500 0 0	£562 10 0		£2500 0 0	£562 10 0

COMPOUNDING FUND

Statement 6

	Investments £ s. d.	Uninvested Capital £ s. d.	Revenue £ s. d.		Investments £ s. d.	Uninvested Capital £ s. d.	Revenue £ s. d.
New Investment		100 0 0		Balances at 1 January 1960	2407 9 7	89 15 0	
Interest transferred to Capital			90 0 8	Compounding Fees received		31 5 0	
Balances at 31 December 1960 ...	2507 9 7	211 0 8		Investment Income			90 0 8
				New Investment	100 0 0		
				Interest transferred to Capital		90 0 8	
				Transfer from General Funds		100 0 0	
	£2507 9 7	£311 0 8	£90 0 8		£2507 9 7	£311 0 8	£90 0 8

TRUST FUNDS

FOULERTON AWARD FUND

Statement 7

	£	s.	d.		£	s.	d.
Award to J. Myers	10	0	0	Balance at 1 January 1960	16	10	10
Balance at 31 December 1960	17	0	6	Investment Income	10	9	8
	£27	0	6		£27	0	6

HENRY STOPES MEMORIAL FUND

Statement 8

	£	s.	d.		£	s.	d.
Balance at 31 December 1960	34	6	8	Balance at 1 January 1960	26	7	2
				Investment Income	7	19	6
	£34	6	8		£34	6	8

G. W. YOUNG GEOLOGICAL INVESTIGATION FUND

Statement 9

	£	s.	d.		£	s.	d.
Grant to W. T. Dean	15	0	0	Balance at 1 January 1960	47	10	11
Balance at 31 December 1960	42	10	11	Income from Public Trustee	10	0	0
	£57	10	11		£57	10	11

TREACHER FUND FOR PLEISTOCENE RESEARCH

Statement 10

	£	s.	d.		£	s.	d.
Grant to H. E. P. Spencer	20	0	0	Treacher Bequest	200	0	0
Balance at 31 December 1960	180	0	0				
	£200	0	0		£200	0	0

Statemen

SCHEDULE OF INVESTMENTS—31 DECEMBER 1960

	AT COST OR PAR VALUE WHEN ACQUIRED £	s.	d.		INVESTM INCOME (GR £
FUNDS FOR SPECIAL PURPOSES:					
Bequest Fund:					
£1379 13 3 3½% War Stock	1141	14	0		48
£100 3% Savings Bonds 1960/70	100	0	0		3
£441 5 4 3% British Transport Guaranteed Stock 1978/88	400	0	0		13
£338 6 4 4¼% British Electricity Guaranteed Stock 1974/79	300	0	0		14
£21 0 10 3% Metropolitan Water Board 'B' Stock	20	0	0		1
£806 9 0 6% Newcastle-on-Tyne Stock 1973/76	860	0	0		48
£1728 3 4 5% Conversion Stock 1971 ...	1600	0	0	New Investment	N
	£4421	14	0		£127
Curry Fund:					
10,000 Shares Currys Ltd. par value ...	2500	0	0		562
Compounding Fund:					
£1217 2 6 3% Irredeemable Nottingham Corporation Stock	1152	13	2		36
£200 3% Savings Bonds 1955/65	200	0	0		6
£170 3½% War Stock	152	14	5		5
£100 3% Savings Bonds 1965/75	100	0	0		3
£300 3½% Treasury Stock 1977/80 ...	282	2	0		10
£224 13 3 4¼% British Electricity Guaranteed Stock 1974/79	200	0	0		9
£177 11 4 6% New Zealand Stock 1976/80	180	0	0		10
£131 4 6 6% Newcastle-on-Tyne Stock 1973/76	140	0	0		7
£108 5% Conversion Stock 1971	100	0	0	New Investment	ni
	£2507	9	7		£90
TRUST FUNDS:					
Foulerton Award Fund:					
£349 7 0 3% British Transport Guaranteed Stock 1978/88	294	11	0		10
Henry Stopes Memorial Fund:					
£265 19 2 3% Newfoundland Stock 1943/63	260	12	9		7

H. M. MONTFORD, *Treasurer*.

J. CARMICHAEL, *Hon. Accoun*

We have compared these Statements with the books and records presented to us and find them to a
We have also verified the Investments and Cash Balances held by the Association at 31 December 1

25 January 1961

F. A. MIDDLEMISS
K. C. FORD } *Audi*

PUBLICATIONS

Publications Committee. The Committee consisted of the seven officers with Messrs. D. F. W. Baden-Powell, D. Curry, S. C. A. Holmes, J. F. Kirkaldy, L. F. Hawkes, R. V. Melville, M. P. Kerney and H. W. Ball.

The Committee met on eight occasions and considered twenty-seven papers, of which seventeen were recommended for publication. Five Field Meeting reports were also recommended for publication.

Proceedings. Part 4 of Volume 70 was published in April. This consisted of 78 pages and 4 plates, making a total of 350 pages and 7 plates for the volume. All four parts of Volume 71 were also published within the year with a total of 463 pages and 11 plates.

Guides. The following Geologists' Association Guides were published in 1960 (making a total of 19 guides so far published): Sheffield, Cardiff, Swansea, Belfast, Central Weald, North-East Scotland.

MEETINGS

Nine Ordinary Meetings were held at which three papers were read and four lectures given. Mr. L. J. Pitt (Vice-President) took the Chair at the January meeting, Professor David Williams (President) at the February meeting and Mr. L. J. Pitt (President) at the remainder, with the exception of the June meeting at which the Chair was taken by Mr. D. Curry (Vice-President).

The thanks of the Association are due to the lecturers, to the authors of papers and to the Geological Society for use of its apartments throughout the year.

The names of the ninety-two members elected in 1960 appeared in Circulars numbered 619–628.

1 *January*. Paper: 'The Cenomanian Limestone and Contiguous Deposits West of Beer, South Devon', by W. E. Smith.

5 *February*. Paper: 'Arctic Volcanoes and Glaciers', by Members* of the 1959 University of London Expedition to Jan Mayen Island, Greenland Sea.

4 *March*. Annual General Meeting followed by Ordinary Meeting.

1 *April*. Lecture: 'Bauxite and Karst Geology in Jamaica', by V. A. Eyles.

6 *May*. Lecture: 'The Channel Tunnel Project', by J. M. Bruckshaw and D. J. Carter.

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* From Birkbeck College: A. T. J. Dollar (Leader and Senior Geologist), F. J. Fitch (Deputy Leader and Geologist), J. Banfield (Geologist), K. H. Brinsmead (Geologist), and from Imperial College: B. Chadwick (Glaciologist), D. J. Kinsman (Glaciologist), P. Smith (Glaciologist), and G. Wright (Glaciologist).

10 *June*. Lecture: 'Recent Sedimentation in The Wash', by Graham Evans.

1 *July*. Paper: 'Coastal Geomorphology in the Light of Carbon-14 Dating', by Rhodes W. Fairbridge (Columbia University, New York).

5 *November*. Ordinary Meeting followed by the Reunion (for list of exhibits see page 154).

2 *December*. Paper: 'Some Problems of Fissure-Eruption', by A. T. J. Dollar and J. E. Guest.

At the January meeting Mr. F. J. W. Holwill and Dr. F. A. Middlemiss were elected auditors of the Association's Accounts for 1959.

The Annual General Meeting was held on 4 March 1960, Professor D. Williams, President, in the Chair.

The Annual Report of the Council (previously circulated) was taken as read. It was proposed by Mr. J. C. Piffe-Phelps and seconded by Mr. C. J. R. Braithwaite and duly carried, ' That the Report of the Council, including the Statement of Accounts, be adopted as the Annual Report of the Association for 1959'.

The President declared the following members duly elected as Officers and Members of the Council in accordance with Rule XIII: *President*, L. J. Pitt, F.G.S.; *Vice-Presidents:* D. Curry, M.A., F.G.S., A. K. Wells, D.SC., F.G.S., C. W. Wright, M.A., F.G.S., L. Hawkes, D.SC., F.R.S., F.G.S., David Williams, D.SC., PH.D., M.I.M.M., F.G.S.; *Treasurer:* H. M. Montford, O.B.E., B.SC., F.G.S.; *Secretaries:* General—F. H. Moore, B.SC., PH.D., F.G.S.; Field Meetings—J. M. Hancock, M.A., PH.D., F.G.S.; Publications Committee—D. V. Ager, B.SC., PH.D., D.I.C., F.G.S.; *Editor:* G. T. Raine, B.SC., A.M.INST.PET., F.G.S.; *Librarian:* A. J. Smith, B.SC., PH.D., F.G.S.; *Twelve other Members of Council:* M. K. Wells, M.SC., PH.D., F.G.S., Mrs. J. N. Carreck, G. F. Elliott, F.G.S., John Sutton, PH.D., A.R.C.S., F.G.S., Mrs. M. Ainsley, F.G.S., Miss M. M. Sweeting, M.A., PH.D., C. F. Tozer, B.SC., PH.D., D.I.C., F.G.S., D. G. A. Whitten, B.SC., A.R.C.S., F.G.S., H. W. Ball, B.SC., PH.D., F.G.S., R. V. Melville, M.SC., M. P. Kerney, B.SC., PH.D., G. H. Dury, M.A., PH.D., F.G.S.

It was proposed by Mr. S. Colley and seconded by Mr. B. W. Collins and duly carried, 'That the best thanks of the Association be given to the retiring President, the Officers, retiring Members of Council and the Auditors'.

On the nomination of the Council the following members were elected as Honorary Members: Charles P. Chatwin, M.SC., F.G.S., Leslie Reginald Cox, O.B.E., M.A., SC.D., F.R.S., F.G.S.

The Foulerton Award and Certificate were presented to Mr. John Myers, B.SC., F.R.G.S., F.G.S., in recognition of his work for the Geologists' Association.

The President said:

> John Myers: Our Association is joined, in an especial way, to the interests and aspirations of the amateur geologist. Geology itself, indeed, has always depended upon dedicated enthusiasm in observation and research, and today the Council gladly recognises the enthusiastic devotion with which you have served Geology and the Association.
>
> In 1948, prompted by the eager response of your Workers' Educational Association classes, you called the Meeting which founded the North Staffordshire Group, and in the following year became its first Chairman. You have since continued to serve the Group as Field Secretary, Committee Member and Deputy Chairman, without interruption. You have directed several Field Meetings, and addressed many Ordinary Meetings of the Group.
>
> As the Group's representative to the Centenary Year Committee, you were responsible for co-ordinating the work for those of the Guides associated with the University towns.
>
> Your skill and keenness as an undergraduate at King's College, London, gained you the award of the Tennant Prize, and your service to the North Staffordshire Field Club, as Secretary of its Geological Section, has already been recognised by the bestowal of the Spanton Medal.
>
> The thoroughness of your personal research in the field led you to discover Ordovician fossils in the supposed Silurian rocks of the east side of Newport Bay, Cardigan. More recently, your identification of *Anthraconauta tenuis* in the Etruria Marl Pit, has led to revision of the previously accepted top of the Upper Coal Measures in North Staffordshire.
>
> As a schoolmaster, in the forefront of the development of Geology as an Advanced Level Course in North Staffordshire, you have been responsible for inspiring many young men to an interest, or a career, in Geology, and the satisfaction and pleasure which they owe to you is reflected by us in making this Foulerton Award today.

The President then delivered his address entitled, 'Genesis of Sulphide Ores'.

It was proposed by Professor J. H. Taylor and seconded by Dr. W. Campbell Smith and duly carried, 'That the best thanks of the Association be accorded the President for his Address'.

FIELD MEETINGS 1960

The Committee consisted of the seven officers with Messrs. J. S. H. Collins, G. H. Dury, F. A. Middlemiss, N. B. Peake, J. G. Samson, D. J. Shearman, Miss M. M. Sweeting and Mr. D. G. A. Whitten.

Four demonstrations, two week-end, six whole day and two half-day Field Meetings were held during the year.

The Easter Field Meeting was based on Hastings. Owing to the sudden eath of Dr. F. Jones the Whitsun Meeting to Charnwood had to be ancelled and was replaced by a study of the Mineralogy of the Mendips. For the summer a joint Field Meeting was held with the British Association to study the Southern Shores of the Irish Sea. A meeting in Northern reland was cancelled from lack of support.

The Association is very grateful to the Directors, Secretaries and all others who helped at these meetings.

The following Field Meetings and Demonstrations took place:

23 *January*. Second of three Demonstrations on Fossil Identification, at the Department of Geology, Imperial College of Science and Technology. Demonstrators, Dr. D. V. Ager (Brachiopods), Dr. J. H. Callomon (Ammonites), Mr. F. J. W. Holwill (Rugose corals) and Mr. C. W. Wright (Asteroids).

20 *February*. Third of three Demonstrations on Fossil Identification, at the Department of Geology, Chelsea College of Science and Technology. Demonstrators, Dr. C. G. Adams (Foraminifera), Mr. D. Curry (Lamellibranchs), Dr. M. P. Kerney (Gastropods) and Mr. L. J. Pitt (Polyzoa).

25 *March*. Visit to the BP Research Centre at Sunbury-on-Thames at the invitation of Mr. F. Gill, Manager of the Exploration Division.

9 *April*. Field Meeting at Folkestone to mark the Centenary of the first Field Excursion of the Association, Folkestone, 9 April 1860. Director, Dr. R. Casey.

15–18 *April*. Easter Field Meeting to study the Hastings Beds Deltas. Director, Professor P. Allen.

1 *May*. Field Meeting on the Lower Tertiaries at Swanscombe. Director, Mr. D. Curry.

8 *May*. Field Meeting on the Upper Chalk east of Brighton. Director, Mr. N. B. Peake.

21 *May*. Field Meeting at Greenhithe, Kent. Director, Dr. M. P. Kerney.

3–6 *June*. Whitsun Field Meeting to study the Mineralogy of the Mendips. Director, Mr. R. F. D. Parkinson.

24–26 *June*. Field Meeting to study Recent Sedimentation in The Wash. Director, Dr. G. Evans.

16 *July*. Field Meeting on the Fuller's Earth, Nutfield. Director, Mr. G. M. Davies.

14 *August*. Field Meeting on the Gault and Weald Clay at Small Dole, Sussex. Director, Mr. R. A. Milbourne.

7–12 *September*. Summer Field Meeting, jointly with the British Association, to study the Southern Shores of the Irish Sea. Organising Director, Mr. G. F. Mitchell; Directors, Dr. A. Farrington, Mr. F. M. Synge and Mr. W. A. Watts.

11 *September*. Field Meeting around Guildford and Godalming. Director, Mr. B. Ainsley.

24 *September*. Field Meeting on the Oxford Clay at Bletchley and Woodham. Director, Dr. J. H. Callomon.

30 *September*–3 *October*. Field Meeting in Mid- and West Cornwall. Director, Mr. C. E. Leese.

10 *December*. First of four Demonstrations showing the range of the common minerals and rocks; minerals found in Igneous and Metamorphic Rocks; held at Chelsea College of Science and Technology. Directors, Mrs. J. A. M. Cohen and Mr. A. J. Barber.

NORTH-EAST LANCASHIRE GROUP

Chairman, D. H. Learoyd, B.SC., F.G.S.; Vice-Chairman, J. Ranson, A.M.I.MIN.E., F.G.S.; Secretary, H. H. Grieve, B.SC.; Committee, Mrs. J. Ranson, B.A., A. Moore, B.SC., E. Procter, S. Westhead, I. A. Williamson, B.SC., F.G.S.

During the year there were thirteen meetings including six Field Meetings.

22 *January*. 'The Geology of Malhamdale', by I. A. Williamson.

19 *February*. 'The Geology of the Duddon Estuary', by W. Grieve.

11 *March*. 'The Geology of the Isle of Wight', by D. H. Learoyd.

8 *April*. 'The Geology of the Zillertal, the Tyrol', by A. Sergent.

30 *April*. Field Meeting: 'The NW. Limb of the Burnley Syncline.' Director, E. Procter.

21 *May*. Field Meeting: 'Deglaciation Phenomena, Stratigraphy and Structural Features to the SW. of Blackburn.' Director, D. H. Learoyd.

18 *June*. Field Meeting: 'The Carboniferous Limestones D_1 to C_2 of the Silverdale–Arnside Coast.' Director, J. Ranson.

9 *July*. Field Meeting: 'D_2 to D_4 of Sykes Anticline in Whitensdale, Bowland Forest.' Director, A. Moore.

27 *August*. Field Meeting: 'The Sabden Shales and Permo-Trias of the Lower River Darwen.' Director, D. H. Learoyd.

24 *September*. Field Meeting: 'The Carboniferous Limestone C_2–D_1 at Beetham, Sandside, and Silverdale Fells.' Director, L. J. Hobbs.

21 *October*. The thirty-seventh Annual General Meeting. Résumés of the Field Meetings were given by the directors.

18 *November*. 'The Burnley Syncline', by E. Procter.

9 *December*. 'Crinoids of the Carboniferous Limestones in the Clitheroe District', by S. Westhead.

Lectures were held in the Science Lecture Theatre of the Blackburn Technical College by courtesy of the Governors and the Blackburn Education Committee.

MIDLAND GROUP

Chairman, W. G. Hardie, B.SC., F.G.S.; Secretary, I. Strachan, B.SC., PH.D., F.G.S.; Committee, Professor F. W. Shotton, M.B.E., SC.D., F.G.S., Miss G. M. Bauer, A. Chater, G. Farrell, J. E. Clements, M.A., J. F. A. Stark, B.SC.

Five indoor meetings and five Field Meetings were held.

6 *February*. Dr. J. D. Lawson on 'The Siluro–Devonian Boundary'.

5 *March*. Dr. M. Gage on 'Glacial Geomorphology of New Zealand'.

23 *April*. Excursion to Barr Beacon. Director, Dr. M. A. Khan.

6–8 *May*. Week-end visit to Long Mountain and Shelve. Director, Dr. Strachan.

18 *June*. Whole-day excursion to Cheltenham. Director, Mr. Farrell.

9 *July*. Excursion to South-East of Birmingham. Director, Mr. Hardie.

17 *September*. Excursion to Napton-on-the-Hill. Director, Professor Shotton.

15 *October*. Address by the Chairman, W. G. Hardie, on 'The Geology and Scenery of the Dorset Coast'.

19 *November*. Professor Shotton on 'I.G.C. Excursions, 1960'.

17 *December*. Professor R. W. Lounsbury (Purdue) on 'The Structure of the Northern Rockies' and Christmas Party.

Thanks are due to Professor Shotton for the use of the Geology Department, Birmingham University.

NORTH STAFFORDSHIRE GROUP

Professor F. Wolverson Cope, D.SC., M.I.MIN.E., F.G.S.

Chairman, C. S. Exley, M.A., D.PHIL., F.G.S.; Vice-Chairman, D. O. Thomas, F.G.S.; Secretary, T. S. Purcell, B.SC., F.G.S.; Treasurer, J. T. Gleaves, B.A.; Field Secretary, T. S. Jones, F.G.S.; Research Secretary, D. O. Thomas, F.G.S.; Committee, M. D. Cox, F.B.O.A., R. R. Harvey, P. S. Keeling, B.SC., A.R.S.M., F.G.S., J. Myers, B.SC., F.R.G.S., F.G.S., J. C. Parrack, B.SC., F.R.G.S.

During the year seven Ordinary and four Field Meetings were held. The Group was again fortunate in holding the Annual Conversazione in the Geology Department of the University College, at Professor Cope's invitation.

14 *January*. Professor S. H. Beaver: 'Minerals and Planning': a geologist looks at quarries from a new angle.

11 *February*. W. G. C. Austin: 'Granites.'

10 *March*. Annual General Meeting. Dr. C. S. Exley: 'The Geology of Cornwall and Devon—a Review. Part I: Sediments and General Stratigraphy.'

8 *May*. Field Meeting: Oxford Area. Director, E. A. Watkin.

12 *May*. Dr. M. H. P. Bott: 'British Gravity Anomalies.'

12 *June*. Field Meeting: Llandrindod Wells and Llanfawr Quarry. Director, J. T. Wattison.

17 *September*. Field Meeting: Cheltenham Area. Director, D. O. Thomas.

13 *October*. Annual Conversazione in the Department of Geology, niversity College of North Staffordshire.

15 *October*. Field Meeting: Edale and Castleton. Director, J. C. Parrack.

10 *November*. Professor F. Wolverson Cope: ‘Some Aspects of Italian eology.’ Coloured slides.

8 *December*. Members’ Annual Photo and Slide Evening.

SOUTH WALES GROUP

The Inaugural Meeting of the South Wales Group was held in Cardiff n 23 January 1960.

Chairman, Dr. D. A. Bassett, B.SC., PH.D., F.G.S.; Vice-Chairmen, rofessor J. G. C. Anderson, M.A., PH.D., D.SC., F.R.S.E., Mr. B. Norris, .SC., Professor F. H. T. Rhodes, PH.D., F.R.I.C.; Secretary, Mr. G. Askey, .A.; Assistant Secretary, Mr. W. S. Morgan, B.SC.; Field Secretary, Mr. . M. Thomas, M.SC.; Treasurer, Mr. R. Parsons, B.A.; Committee, Dr. . R. K. Blundell, B.SC., Miss K. Goskar, M.SC., Miss Gillian Groome, .SC., Mr. B. Isaac, B.SC., Miss G. Meara, B.SC., Mr. T. R. Owen, M.SC., .G.S., Miss M. I. Roberts, B.SC., Mr. D. Thomas, B.SC.

During the year seven Ordinary Meetings and four Field Meetings were eld. The Ordinary Meetings were held alternately at Swansea and ardiff at the University Colleges but the symposium on the Arctic and ntarctic and the talk to the Junior Members were held in the National useum of Wales.

23 *January*. ‘Prospecting for Minerals’, by Professor D. Williams.

13 *February*. ‘The Geology of South-East Wales, with special reference o the Cardiff Area’, by Professor J. G. C. Anderson.

3 *March*. ‘The Rocky Mountains’, by Professor F. H. T. Rhodes.

5 *April*. First Annual General Meeting. Chairman’s Address: ‘The istory of Lower Palaeozoic Research in Wales, with special reference to he Cambrian rocks of Merionethshire.’

9 *April*. Field Meeting: ‘The Vale of Glamorgan.’ Mr. Trevor M. homas.

23 *April*. Field Meeting: ‘The Head of the Vale of Neath.’ Mr. T. R. wen.

7 *May*. Field Meeting. ‘Beachley–Aust and Chepstow Areas.’ Professor . G. C. Anderson.

21 *May*. Field Meeting: ‘Llandilo–Llandovery Area.’ Professor F. H. T. hodes.

8 *October*. ‘Sedimentation in Geosynclinal Troughs.’ Professor Alan ood.

5 *November*. Symposium: Arctic and Antarctica. Geology. ‘Some spects of the Geology of the Arctic’, by Dr. J. W. Cowie. ‘An Introduction

to the Geology of Antarctica', by Dr. D. H. Maling. 'The Geology of Graham Land and the Scotia Arc', by Dr. P. R. Hooper.

8 *December*. 'Geological Holidays in Scotland.' A talk to the Junior Members by the Chairman.

10 *December*. 'The Geological Approach to Historic Buildings', by Dr. F. J. North.

JAMAICA GROUP

The Jamaica Group was dissolved at the Annual General Meeting held in December 1959. A new Geological Society of Jamaica was formed.

REUNION

The Annual Reunion was held at the Chelsea College of Science and Technology. It was attended by nearly 300 members and friends. Thanks are due to the Principal and Staff of the College and the members concerned.

LIST OF EXHIBITS

British Museum (Natural History)

Plaster cast of Tarbosaurus, a giant carnivorous Dinosaur which was presented to The Royal Society on the occasion of their tercentenary by the U.S.S.R.

Geological Survey and Museum

(*a*) Recent publications.
(*b*) Geological photographs.
(*c*) Photographs of the effects of the Chilean earthquakes, 1960.
(*d*) New 1-inch to 1-mile Mendip Hills, Bristol, and Forest of Dean model.

Ager, D. V.

Simple drawing apparatus for serially ground fossils.

Andrews, Miss E. W.

Details of a section in the Woolwich Beds exposed during the construction of a reservoir at Farnborough, Kent.

Anthony, H. D.

Geology and Scenery:
(*a*) The Staddon Grits.
(*b*) Colour transparencies of the 1960 Easter Field Meeting.

Bennett, W. H.

Some fossils from North America.

Best, A. C. G.

Pebbles from sea-bottom in 700 fathoms on the Continental slope due west of Little Sole Bank in neighbourhood 48° 30′ N., 10° 15′ W. Obtained on cruise of Marine Biological Association's Research Vessel *Sarsia*.

Blezard, R. G.

A lithological comparison of zones of Upper Maestrichtian and Lower Danian in Holland and Denmark with identical foraminiferal fauna and identical orthogenetic stages.

Bottley, Mr. and Mrs. E. P.

(*a*) British fluorspars and associated minerals.
(*b*) Fine Australian opal and opalised fossils—recent finds.
(*c*) Drepanaspis. New discoveries made in 1960 in the Lower Devonian slates of S. Germany.
(*d*) New series of prehistoric animal models for schools.
(*e*) New specimens of fine minerals.
(*f*) New models of twinning in felspars.
(*g*) Models of the Weald and models of hypothetical Field areas for teaching field surveying.
(*h*) A new low-priced fluorescent lamp for schools.

Cambridge, P.

North American Dinosaurs.

Carreck, Mrs. J. N.

(*a*) The Association's photograph albums prepared by Miss M. S. Johnston and others.
(*b*) Photographs donated during 1960, and the new album.

Carreck, J. N.

(*a*) William Smith's geological map of Kent, 1821.
(*b*) Archive material formerly owned by the late Charles Davies Sherborn, including letters from Clement Reid, J. W. Hulke and W. Boyd Dawkins.
(*c*) A new discovery of Reindeer and other mammals in the Buried Channel of the Medway near Maidstone.

Casey, R.

Palaeozoic or pre-palaeozoic boulders from the Lower Greensand of Folkestone.

Collins, J.

The stratigraphical distribution of some Gault decapod Crustacea.

Curry, D.

Venericor planicosta.

Cutrock Engineering Co. Ltd.

Projector polaroid unit. Mark II Unicutta thin section cutting and grinding machine. L.M.S. 50 lapping machine. C.H.P. 100 hotplate. Abrasives. Hammers. Chisels, Augers. C.R.S. 100 rock-splitting and crushing machine. Stereometer. Dr. Dollar's integrating and point counting micrometer. Dr. Dollar's clinometer. Croft grinder. Diamond lapping unit.

Dutt, P. K. & Co. Ltd. (R. H. Freere)
Equipment for Geology.

Gibson, I. L., D. J. J. Kinsman and G. P. L. Walker
Folding in rhyolite lavas from E. Iceland.
Chalcedony from amygdales in Tertiary basalts of E. Iceland.
Geological photographs, E. Iceland.

Ivimey-Cook, H.
A new Triassic alga.
Colour transparencies taken at Field Meetings in 1960.

King's College, London
1955 Eruption at Kilauea, Hawaii Islands. Film.

Kingston Technical College, Geology Section
More new Scottish Rocks.

Large, Nicholas
The Rhaetic exposures at Aust and Westbury Garden Cliffs, Glos.

London Natural History Society
Easter Field Meetings of the Geological Section, 1949–60.

London University Geological Expedition to Skidadlur, 1960
Specimens, photographs and diagrams to illustrate the scope of a month's research in the Skidadlur region in Northern Iceland.

Marston, A. T.
Chert Acheulean hand-axes from Barnfield Pit, Swanscombe, Kent.

Milbourne, R. A.
The Stratigraphy and Ammonite succession of the *lautus* zone of the Lower Gault.

Pitt, L. J.
Crag polyzoa.

Ramsey, D. M. and B. A. Sturt
Igneous and Metamorphic Rocks from Sørøy, Northern Norway.

Ray, Marion
Paintings of minerals.

Rayner & Keeler Ltd. (C. H. Smith)
(*a*) Fluorescent minerals.
(*b*) Fine mineral specimens.
(*c*) Interesting fossils.
(*d*) Ultra-violet light sources.

Ross, F. Stenhouse
Material from the Cephalopod Bed, Wooton-under-Edge, Glos., and minerals collected in 1960.

Royal School of Mines (Mining Geology Department)
Geological coloured photographs taken underground in the Lake District and Pennines orefield areas.

Smith, W. E.

Rocks, minerals and fossils collected in Denmark during the International Congress, 1960.

Spencer, H. E. P.

Crag mammalian remains.

1960 University of London Expedition to Jan Mayen Island, Greenland Sea

Rocks collected by A. T. J. Dollar and J. E. Guest during the Arctic summer.

Venables, E. M.

Fossil insects from the London Clay of Bognor Regis collected in 1960.

Walker, G. P. L.

Rocks and minerals from southern Norway and Sweden.

Wooldridge, S. W. and H. C. K. Henderson

Map showing the surface of Essex and Suffolk.

Wooldridge, S. W. and E. H. Brown

The Plio-Pleistocene shore lines of southern Britain (Map illustrating Paper to Section 'C' at Cardiff meeting of the British Association).

Wright, C. W.

Recent acquisitions.

Wyley, J. F.

(*a*) Vanadiferous and Uraniferous nodules from the Permian at Budleigh Salterton, some showing an undescribed vanadium mineral comparable with Pascoite.

(*b*) Anakieite. A new Hornblende–Pyroxene–Garnet Rock from Mount Anakie, Australia. Named and sent by J. Hollis, Geelong Grammar School.

(*c*) A porcellaneous variety of Basaluminite from Newhaven, Sussex.

THE LIBRARY

The Library Committee consisted of the President, Mr. L. J. Pitt, J. W. Scott, B.A., A.L.S., Librarian of University College, London, and A. J. Smith (Librarian).

The following institutions were added to the Exchange List in 1960:

The Geological Institute, University of Athens, Greece.

The Geological Survey of Tanganyika.

The Gorki Scientific Library of Leningrad, U.S.S.R.

The State Library for Scientific and Technical Books, Moscow, U.S.S.R.

DONATIONS TO THE LIBRARY

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OBITUARY NOTICES, 1961

FRANCIS HEREWARD EDMUNDS, who died at Westhumble, Surrey, on 9 April 1960, will be sadly missed by officers and members of the Association, which he joined in 1922. For some three years he had been retired from the Geological Survey of Great Britain, where the greater part of his life's service to geology was carried out. It came as a shock to his friends and former colleagues to learn that his strong Wiltshire constitution had unhappily failed to see him through a second bout of serious illness.

Edmunds was born at Mere in 1893, one of a large family who have retained the interests of countrymen while leading busy professional lives. For him the practical science of geology became an absorbing study when, after education at Gillingham Grammar School and serving in the Tank Corps during the First World War, he was able to go up to Emmanuel College, Cambridge. In Part II of the Natural Sciences Tripos he obtained first-class honours in Geology in 1922.

The Geological Survey, which he joined in London the same year, saw Edmunds carrying out detailed mapping in Surrey and elsewhere and later preparing exhibits for the new Museum at South Kensington. In the popular *British Regional Geology* series of handbooks he was sole author of *The Wealden District*. He was part author of several explanatory memoirs, notably those on the Guildford and Reigate Districts. Among various aspects of official advisory work Edmunds specialised on civil engineering and underground water-supplies. The latter work bore fruit in a number of contributions to scientific journals, particularly the *Transactions of the South-Eastern Union* (1934) and the *Journal of the Institution of Water Engineers* (1941), as well as in Geological Survey memoirs such as *The Wells and Springs of Sussex*; the former found outlet in his own book *Geology and Ourselves*, which includes treatment of geology in everyday life that has a deservedly popular appeal. While District Geologist of South-Eastern England, from 1945 to 1957, Edmunds maintained his enthusiasm for field work while supervising new programmes of six-inch mapping and the re-issue of many New Series one-inch sheets.

The Association is indebted to Edmunds for leading various excursions in the Guildford and Dorking districts, at West Hoathly and in the Mere district of Wilts. He published in the PROCEEDINGS, with H. G. Dines, an account of the *Tectonic Structure of the Hog's Back* (1927) and, in co-operation with R. J. Schaffer, an authoritative description of *Portland Stone: its Geology and Properties as a Building Stone* (1932). He wrote a useful *Contribution on Physiography of the Mere District* in 1938.

Edmunds was president of the South-Eastern Union of Scientific Societies in 1949 and took a prominent part in the Geology Section at the British Association meetings of 1954 and 1955.

The friendly guidance, encouragement and advice, which he so willingly put at the disposal of innumerable students, research workers, amateur geologists or naturalists and interested laymen, revealed a generous desire to impart geological knowledge which Edmunds was always pleased to implement by lecture, field demonstration or discussion, no matter how modest the circumstances or how relatively inexperienced the colleague or visitor.

If Edmunds was a practical man in geology, those who had the pleasure of visiting him at his home near Box Hill will recall his no less practical achievements in home and garden. He was also one who had read widely and both as a performer and listener had acquired a devoted interest in musical affairs. His services in recent years both to the Leith Hill Music Festival and as secretary during the distinguished activities of the Mickleham and Westhumble Music Festival fully deserved the recognition they received. His wife and daughter will also realise how appreciatively his energetic personality will long be recalled in geological circles.

S. C. A. H.

GERTRUDE LILIAN ELLES, born at Wimbledon, Surrey, on 8 October 1872, died on 18 November 1960 in a nursing home in Helensburgh shortly after leaving Cambridge. She had joined the Association in 1901.

Although Miss Elles had spent nearly the whole of her life in Cambridge it was seemly that she should have died in Scotland for throughout her life she was an ardent Scot returning to her native land whenever the opportunity presented itself.

She entered Newnham College, Cambridge, as a scholar in 1891, taking a First Class in Geology in the Natural Sciences Tripos Part II in 1895.

In the old Woodwardian Museum, which was housed in the Old Schools next to the Senate House, she came under the influence of McKenny Hughes and his brilliant team of teachers, particularly Marr. It was natural that she should be guided to an interest in Lower Palaeozoic rocks and their fauna. It thus came about that she started work on the graptolites and when a research studentship from Newnham College was awarded to her she availed herself of the opportunity to visit Sweden and to study the beautiful material found in that country, where she also had the advantage of working at Lund under Törnquist.

On returning to England it was inevitable that anyone working on graptolites should come into close contact with Lapworth and thus she and Miss E. M. R. Wood (Dame Ethel Shakespear) began the preparation of the monumental monograph on the British Graptolites under the editorship of Professor Lapworth.

The names Elles and Wood will long be linked together in this great standard work which forms the basis of all work on British graptolites for many years to come.

In 1922 Miss Elles published an outstanding essay in the PROCEEDINGS in which she brought together her general conclusions regarding the evolution and classification of the graptolites from her life-long study of the group.

Although she will be remembered mainly as an authority on the graptolites she was deeply interested in the general problems of Lower Palaeozoic stratigraphy and published a number of papers on the subject, the best known being those in the Ludlow area, Bala and Builth. In these she showed that she was, at heart, a field geologist in the tradition of the Cambridge School. She always looked at a fauna as a community of animals living together rather than considering them as isolated 'specimens' on a laboratory bench. This aspect is well seen in her Presidential Address to Section C of the British Association in 1923. It might have been expected that she would have dealt with some aspect of graptolite evolution or distribution, she chose however to discuss stratigraphical palaeontology illustrated by modern faunal assemblages and evolution of various fossil groups.

An ardent Scot, she always yearned for the Highlands and whenever free to do so spent her holidays there, fishing being one of her main recreations. It was thus that she became deeply interested in the structures and origins of her beloved mountains, but it shows the breadth of her geological interests that comparatively late in life she, with Professor Tilley, entered into their detailed study and produced an important paper on the problems involved.

It may well be that it was a study of Scotland that was the motive force rather than an overwhelming interest in metamorphism and structure.

Apart from her geological work she will be remembered as a great teacher and an outstanding member of Newnham College. To this College she was deeply devoted having been college lecturer, Fellow, Vice-Principal and finally Honorary Fellow. In all these posts she took her full share of administration. But it was in her relationship with the junior members of both her own and those of Girton College that her influence was most marked. As a teacher her enthusiasm and high standards attracted many to the study of geology and taught them what was expected of a serious student of science. Her influence was not confined to the women students and many men undergraduates will remember the help and inspiration they received from her.

In the year she retired from her College posts she was appointed to a University Readership, the first woman to hold this position. She took the degree of D.Sc. (Dublin) in 1907 and Sc.D. (Cantab.) in 1949.

For many years she was an active member of the British Red Cross Society and from 1915 to 1918 was Commandant of a small convalescent home for soldiers just outside the College gates, being awarded the M.B.E. for her services.

Until prevented by increasing deafness she took great pleasure in music of which she had considerable knowledge and great appreciation and was a regular attendant at King's College Chapel; those who attended the beautiful memorial serice held there will have realised how fitting it was.

It was typical that when she retired to 21 Barton Road, Cambridge, she called her house 'Caladh', the old Highland name for the inner harbour where boats, too old for going to sea in all weathers, are laid up. It was not in Cambridge, however, but in her native Scotland that she found her true 'Caladh' only a few days before her death.

W. B. R. K.

A very full obituary notice of the late Mr. S. H. Warren appeared in the PROCEEDINGS of the Geological Society for 7 October 1958, but the pages of our PROCEEDINGS should not pass without a note of one of our Presidents and the author of a number of important investigations among the Crag Deposits of East Anglia following up certain lines of human archaeology. I knew Warren well at one period of his membership, stayed with him at Loughton after his wife's death and so saw much of his collections which he had arranged in open frames in one of his rooms. Within the last sixty years when the exploitation of loam in Middle Deal for brickmaking had ceased but was still carried on at Sholden and Cold-Blow, Walmer, Warren had worked on these sites. Articles from them formed part of his collections.

Warren joined the Association in 1897 and was elected an Honorary Member in 1933. He was one of the Association's Managing Trustees for many years, President 1922–4 and Treasurer 1914–21. He was the recipient of the Henry Stopes Memorial Medal in 1949 and delivered the second Memorial Lecture entitled 'The Clacton Flint Industry: A new interpretation', on 5 May 1950, in which it was suggested that the Clacton industry was of very early tradition—a modified survival of the early pebble-tool industries of Asia and Africa which spread into Western Europe as a major glaciation passed away. His first Presidential Address (1923) was on 'Sub-soil Pressure-Flaking', and the second (1924), on 'Pleistocene Classifications'.

I cannot close this note by one of his contemporaries whose knowledge, however, goes back to older stalwarts who built up our 'amateur' Association, without referring to Warren's many Kentish finds and also to 'The Clacton Spear' which Oakley refers to as 'the oldest known wooden artifact in the world', and figures as number five on page fifteen of the British Museum Guide, *Man the Tool-Maker*, 1950.

W. P. D. S.

An Account of the Work on Polyzoa

by L. J. PITT

Presidential Address read 3 March 1961

THE SEVERAL GROUPS making up the sessile marine animals were completely misunderstood until a little over two hundred years ago. Early in his life Linnaeus expressed the general view of his time in the definition 'stones grow, vegetables grow and live, animals grow, live and feel'. No one had observed these sessile groups sufficiently closely to determine their modes of reproduction and the short, free swimming larval stage of the Polyzoa, Hydrozoa and Corals had not been noticed. They were quite naturally regarded as plants.

This is a brief account of some of the many workers who endeavoured to unravel the misunderstandings, the opposition they met and the measure of success they achieved in connection with the Polyzoa, but in order to follow the story through it will be necessary to mention other groups also.

As a preface I should like to quote to you from George Johnston, who, in 1838, published a work on Zoophytes and endeavoured to give a brief account of the authors of the species he mentioned. He wrote: 'This has been a pleasing enquiry. Smit with the beauty—real or fancied—of the objects of his study, a curiosity is naturally awakened to discover the name and degree of the person who first deemed it worthy of his examination and participated in our pleasure, for we conclude assuredly that he who had taken the trouble to record the name and treasure up the object, was one of like mind, and imbued with much of the same affections and dispositions as ourselves. Some of them were found to be men of renown, others, in whom I felt a deeper sympathy, are now forgotten, their name and their labours swallowed up in the higher and more enduring reputation of those whom they were honoured to assist and delighted to serve.'

If there were no living Polyzoa it is certain that we could not understand the fossil species nor classify them properly. In the Hydrozoa and Corals there is no alimentary canal, the mouth opens into a sac, and it is the possession of this complete alimentary canal which distinguishes the Polyzoa as a separate group. This is of course a soft part of the animal and never found fossil.

The Polyzoa have been variously described as colonial animals or compound animals, and one worker, Levinsen, called them 'aggregate animals'. The whole colony is termed the zoarium, a name introduced by Busk, and is made up of a number of tubes or chambers, termed the zooecia, which are the calcified body walls of the soft individual animals. All the normal animals have an individual life and can survive if other parts of the zoarium are damaged, yet in another sense the zoarium itself is a living whole, the normal animals supplying nourishment to the modified individuals. There is usually polymorphism in the Polyzoa, some individual animals being modified for special functions.

The individual zooecium is very small, 0.1 to 0.2 mm. in diameter and 0.4 to 1 mm. in length being the usual order of measurement in Cyclostomes. Consequently it is not difficult to see that progress in understanding Polyzoa was linked with the development of the microscope.

The first published figure and description of a Polyzoan is in a general work on Natural History by the Neapolitan chemist Imperato, published in 1599. It is shown encrusting a seaweed found below low-water mark off the Italian coast and commonly washed up after storms. In his day the incrustation was scraped off, crushed and sold under the name of 'adarce'. This was used for cleaning teeth and was probably identical with the 'adarce' of classical writers, though it is obvious from Imperato's work that there were other sources of the material. In his Presidential Address to the Linnean Society in 1929 the late Sir Sidney Harmer dealt at some length with Imperato's figure and description, identifying the seaweed as *Posidonia oceanica* and the Polyzoan which Imperato had named '*Poro aguino*' as the modern *Calpensia nobilis*. Elsewhere in his work Imperato mentions names still in use for Polyzoan genera today—Retepora and Frondipora.

Exactly a hundred years later Edward Lhuyd of Oxford published his classic work on fossils. Fortunately his collection has been preserved in the Museum of the History of Science at Oxford and among his specimens there are at least two Polyzoa from the Lower Greensand Sponge Gravels of Coxwell, near Faringdon. One of these he figured as *Porus minimus* but unfortunately both specimens are too worn for certain identification. The quarries in this area have been worked for a very long time and yield a rich Polyzoan fauna in addition to the Sponges for which they are best known. We shall come back to this area later in our story.

Imperato was a little too early to arouse interest in the minute, as was Lhuyd to foster any enthusiasm for geology. Progress is the result of team work, discussion and criticism, and they were both lone workers. Even in Lhuyd's day microscopes were poor and the time was not ripe for the study of small things.

In 1729 a French physician-botanist, Peyssonnel, discovered the animal nature of some of the marine organisms which had been regarded as plants and in all probability he worked on Hydrozoa. He was not a member of the French Academy and so could not announce his discovery in person, but he had a friend, Réamur, who was, and to him he entrusted his paper. Réamur promised to read it for him but was not at all happy about such a revolutionary idea. He thought the paper would bring ridicule on his friend, so he kept his promise and read it without naming the author. Then he went on to speak against it. Naturally it was not published and shortly after it disappeared. Peyssonnel was convinced he was right but before he could do much more work he received an official appointment to the island of Guadeloupe. There he was away from Europe and contact with other work going on but he himself was able to carry on his researches.

Following Peyssonnel, Abraham Trembley had been carrying out some experiments on the mode of reproduction of some freshwater Hydrozoa and Polyzoa, and the publication of his results recalled to Réamur the unfortunate paper of his old friend Peyssonnel. Réamur suggested to another French naturalist, Bernard de Jussieu, that Trembley's experiments should be repeated on marine material, and asked him to collect and make further observations in the South of France. This de Jussieu did and was able to figure the living polypides. He in his turn presented a memoir to the French Academy in 1742, but its reception was lukewarm and did not carry conviction. Peyssonnel heard of this and decided he would make another attempt to advance his claim to the discovery of the animal nature of these so-called plants. He prepared a second paper and forwarded it to The Royal Society in 1751. It was ridiculed by Dr. Parsons, a well-known naturalist of the time, who considered them purely vegetable growths, the animals observed by Peyssonnel being but accidental settlers! Although neither of Peyssonnel's papers survive some extracts have been quoted by other writers which show rather fanciful ideas, and these may well have been the cause of some ridicule.

Later, in 1753, the microscopist, Henry Baker, published his *Employment of the Microscope*, in which he supported a theory that these growths were neither plant nor animal but wholly mineral, like the crystal growths he had seen in his studies.

Independently of Peyssonnel another worker was in the field who was to carry conviction. William Ellis was a London businessman, but a native of Ireland. He was an amateur botanist of some repute and a friend of Linnaeus. In the autumn of 1751 he received two collections of sea plants and corallines, one sent to him from Anglesey, the other from Dublin. He decided to apply to them a method recommended to him by a well-known German naturalist, Büttner. After soaking in water to expand them he transferred them to paper and finally mounted them on white card to make a landscape picture. His friend Dr. Hales saw it and suggested he should make another for the amusement of some small members of the Royal Family, so with the assistance of his friends, Ellis made a collection of all the varieties of seaweed they could obtain round the coasts of the British Isles. The offering was duly prepared and the Dowager Duchess of Wales accepted it graciously, though there is no record of the reactions of the Royal children! However, the great variety of material which passed through his hands led Ellis to work over the remainder and endeavour to name it, using the standard work of his day, Ray's *Synopsis of British*

Plants. He found it necessary to use a microscope for this work and had an instrument made to his own specification. Later, Bate, a London maker of microscopes, marketed a modified version of this and we know what it was like, for one is in the collection of the Royal Microscopical Society, catalogued as 'a modification by Bate of Ellis' Aquatic microscope'. It was a simple type of dissecting stand. Ellis soon came to the conclusion that many of the so-called plants in this collection were more like animals than vegetables. To make certain he decided that he must see living material, and in August 1752 he went to Sheppey, taking with him a painter to make proper drawings of the things he found. On this occasion he used a microscope made by Cuff and modified by himself, a conventional tube instrument mounted on a box. Again we know what it was like for Ellis figured it at the end of his *An Essay towards the Natural History of the Corallines*. His next effort was a visit to the seaside village of Brighthelmstone in Sussex—the modern Brighton—in June 1754. On this occasion he took with him a botanical illustrator named Ehret, and they were able to see the polypides moving and make drawings. They also observed the ovicells of some and Ellis understood them as such. He gave an account of this visit to The Royal Society and showed Ehret's drawings. In August of the same year he took a visitor, Dr. Oeder, Physician and Botanist to the King of Denmark, on a journey to Whitstable, where he enlisted the fishermen to collect for them. Ellis now wanted to preserve specimens with the polypides extended, so as the material was brought to them he put it into buckets of sea-water. When he could see the tentacles extended he snatched the specimen from the sea-water and plunged it into brandy. Apparently the cost was considered worth while, for he succeeded in getting the state of preservation desired. He set out the results of his researches and his conclusions in *An Essay towards the Natural History of the Corallines* published in 1755 The figures are extremely good and the work was the first real advance in establishing the true nature of the group which Linnaeus unfortunately named the Zoophyta. Ellis had correspondence with Linnaeus, who was dubious about the animal nature of sessile organisms, but he thought too highly of Ellis's work and reputation to contradict him. He conceded that they were a half-way house between animal and plant, hence the unfortunate name.

When Büttner, who had passed on the technique for mounting seaweeds, heard of Ellis's work he tried to steal the credit for himself, claiming that he had sent a paper to The Royal Society which Ellis had copied! Ellis applied to the secretaries of the Society and received from them a certificate stating that they had never had a paper from Büttner. No more was heard of Büttner's claim. Ellis seems to have had both the time and money to pursue his scientific work but one record of his life states that he was not a very successful businessman until later in his career. He was granted the agency for West Florida in 1764 and that for Dominica in 1770, and these appointments not only put him on his feet but also gave him opportunities to import seeds and so further his botanical studies. Perhaps he gave too much time to the things he thought worth while and too little to his business. Ellis died in 1776, and his friend Dr. Solander edited a volume of his notes and plates on the Zoophytes but himself died before it was published in 1786. Ellis & Solander's work dealt largely with corals, although Linnaeus had already decided in 1767 that the Corals should be separated from the Zoophyta. Most naturalists agreed with this, but many still followed the easy course of considering the Corals with it.

The next big step forward was in 1827 when Dr. Robert Grant described the animal of the common British Polyzoan genus *Flustra*. Grant was born in Edinburgh in 1793 and graduated as Doctor of Medicine in 1814. He chose species of *Flustra* which were abundant on the east coast of Scotland and could be collected and examined at all seasons of the year. He succeeded in rearing the larvae on watch-glasses in an aquarium and gave a very good account of their metamorphosis and the early growth-stages of the young zoarium. His observations were careful and detailed, but he missed the one feature possessed by *Flustra* and all the other Polyzoa but not found in the Hydrozoa—the complete alimentary canal. The following year he was appointed Professor of Zoology and Comparative Anatomy at University College, London, where he remained until his death in 1874. During this time he gave an annual lecture on Palaeontology.

While Grant was making his observations on *Flustra*, the man who was to describe the Polyzoa effectually and designate them as a separate class was working independently on the genus then known as *Sertularia*, and found that the species ascribed to the genus could be divided into two distinct groups with completely different types of polypide. John Vaughan Thompson was born on 17 November 1779 and as a young man lived at Berwick-on-Tweed. At the age of twenty he was an army staff surgeon serving in the West Indies, becoming a full surgeon in 1803. During his stay there he took up the study of Zoology and wrote one or two papers. He returned to England for a few years and from 1812 to 1816 saw service in Madagascar and Mauritius. On his return he left the army to take up the post of district medical inspector at Cork. His study of *Sertularia* commenced in 1820 but he did not publish the very lucid paper setting out his conclusions until December 1830. For the first time the polypide of Polyzoa was correctly described. Thompson proposed that the species of *Sertularia* which were of higher organisation than the Hydrozoa, i.e. those with complete alimentary canals, should be transferred to a new genus, *Vesicularia*. Further, he placed them in a new Class, the Polyzoa. His paper was published in the *Zoological Researches*, in Cork, and it is probable that Ehrenberg, working independently of Thompson, never saw the paper or knew of Thompson's work when he made the same discovery, published his conclusions in 1831 and proposed the name Bryozoa.

The two names have been used ever since, Polyzoa mainly in the British Commonwealth and Bryozoa on the Continent and in the States. From time to time arguments have raged on their merits and I do not propose to comment on them, but I should like to make two observations—Thompson's designation of Polyzoa put them in a new class, Ehrenberg left his Bryozoa in the Phytozoa. Also, the name Bryozoa, 'moss-like animals', savours of the 'plant-like animals' of the Zoophyta and is not an apt description of the whole group.

Just before Thompson published he was promoted to Inspector-General of Irish Hospitals, a post which he held for five years before going to Australia in charge of medical work in the convict settlements. He died in New South Wales on 21 January 1847. His greater responsibilities seem to have prevented him from doing much more zoological work. The *Dictionary of National Biography* makes the comment on Thompson that 'no naturalist has ever written so little and that so good'.

Although naturalists continued to speak of the Zoophytes for another twenty years, Thompson's work was a major step in clearing up the misunderstandings regarding this heterogeneous group and from his time the work on both recent and fossil Polyzoa went ahead rapidly.

Up to Thompson's time a little work had been done on fossil Polyzoa, under the name of Zoophytes or Corallines. The standard work on palaeontology in this country in the early part of the nineteenth century was Parkinson's *Organic Remains of a Former World*, published between 1804 and 1811. The second volume, dealing with the Zoophytes, came out in 1808 and some of the figures are undoubtedly of Polyzoa, though drawn on too small a scale to show useful detail. His localities are very vague—'from Berkshire' or 'from Switzerland' being all that he considered necessary. James Parkinson was a doctor who had had a practice in Hoxton at least from 1785 and he was still there when the *Organic Remains* was published. He was a founder member of the Geological Society and in a paper read at an early meeting showed a lively appreciation of the importance of William Smith's work on stratigraphy. He was also a member of one or two politically suspect bodies, and had correspondence with one of the men accused in the 'Pop Gun' plot to assassinate George III by shooting a poisoned dart at him in the theatre. No proceedings were taken against Parkinson, but he was examined under oath on account of his connection with the accused.

In France, Lamouroux described many Polyzoa from the Jurassic of Normandy during the first quarter of the century and some early work was done by de Blainville. Henri Milne-Edwards published a paper on the Eschara in 1836 and later joined with Jules Haime in work on the Palaeozoic Polyzoa, which they published in 1848 and 1849. (The British name of a French palaeontologist is interesting—his father was a Jamaican planter who left the island owing to political unrest and settled in Bruges, where Henri was born. The young Belgian then went to Paris to study and took French citizenship!)

In the decade following Thompson's work here and Ehrenberg's on the Continent, Karl von Hagenow published his work on the Rügen Chalk Polyzoa and George A. Goldfuss and F. A. Römer both dealt with other German material. Hardouin Michelin published in France between 1841 and 1847 and A. E. von Reuss's work on the Bohemian Cretaceous Polyzoa appeared in 1845 and 1846. Reuss later dealt with Tertiary Polyzoa and did the first work on the Austrian Miocene. In this country William Lonsdale did some early work but later turned his attention to American Polyzoa. Alcide d'Orbigny's monumental works covering the whole field of palaeontology were published between 1849 and 1854 and included much work on Polyzoa.

In this country the attention was on Recent Polyzoa rather than on fossil in the period immediately after Thompson's work. An early amateur worker, George Johnston, published his *British Zoophytes* in 1838, stating that his purpose was to draw together in one volume all the knowledge of the Zoophytes. He did, however, propose the division of the Polyzoa into two orders, Freshwater and Marine, an arrangement which foreshadowed Allman's later work. Johnston's wife assisted him by drawing the figures and even did some of the engraving. He devoted considerable space to the early controversies on the animal nature of the Zoophytes. He was born at Simprin in Berwickshire on 20 July 1797, and was educated at Kelso and Berwick Grammar Schools before graduating as a Doctor of Medicine at Edinburgh in 1819. He was a man of tremendous energy, three times mayor of Berwick, at one time editor of the *Annals and Magazine of Natural History*, and the author of about ninety papers. He prefaced his work on Zoophytes, from which I have already quoted a passage, with a word of praise for the many busy fellow doctors who had found time to help him in his work, and made a comment on the use of time which is worth repeating—'Idleness has no leisure'.

David Landsborough, a fellow Scot, was a slightly older man than Johnston. He was born at Dalry in Galloway on 11 August 1779, educated at Dumfries Academy and Edinburgh University, and entered the ministry of the Church of Scotland in 1808. His industry was staggering, even in comparison with Johnston, and he was the first populariser of the Zoophytes. He wished to keep up his scholarship, so, in addition to his work as a minister, he set himself the discipline of a daily reading of some Hebrew, Greek, Latin, French or Italian and then added the study of botany! Soon he widened his studies in natural history, and the discovery of a new species of Polyzoan brought him in contact with Johnston. When the Scottish Church split in 1843 he joined the Free Kirk. He left a church with a stipend of £350 a year to minister at Saltcoats for £120, and the change left him without his much-prized garden. This made him turn his attention to the seashore. His church and its school needed funds, so he set his children to collect seaweeds, which he mounted into sets, named and sold. About the same time, and probably to augment his own income, he contributed some articles on natural history to various periodicals. These attracted attention and he was asked to write a popular book on British Seaweeds which was so successful that the publishers asked for a popular work on Zoophytes. The *Popular History of the British Zoophytes or Corallines* came out in 1852. Readers of the introduction must have turned back to the title-page several times to make sure they were reading a work on natural history and not a book of Landsborough's sermons. None the less, this first popular work on Zoophytes was a fine effort and well illustrated, though the credit for the figures partly belongs to Busk, who we shall hear about later. These two popular works must have helped Landsborough financially, for shortly after, at the age of seventy-three, he made a journey to Gibraltar, Tangier and other places in the Western Mediterranean. Returning home he found the challenge of a cholera outbreak in his district and met it by continuing his sick visitation, bringing what comfort he could to the sufferers. As a result he contracted the disease and died on 12 September 1854. In his life he discovered some seventy or so species new to Scotland and earned the name 'the Gilbert White of Ardrossan'. It was Landsborough who recorded the outsize specimen of *Membranipora* [*Flustra*] *membranacea* encrusting a seaweed, eight inches wide and about five feet long, which he estimated consisted of over two million zooecia.

About the middle of the century an amateur geologist turned his attention to the Lower Greensand of Faringdon, the locality from which Lhuyd had collected one hundred and fifty years earlier. Daniel Sharpe made the first attempt to use the Polyzoa stratigraphically, and unfortunately failed. The Sponge Gravels of Faringdon were not understood and their Lower Cretaceous age had not been determined. Sharpe found in d'Orbigny's works many species of Cretaceous Polyzoa with a superficial resemblance to those he had collected at Faringdon (the volume of the Terrains Crétacés dealing with Polyzoa was issued in 1853). He named most of his material from these figures and concluded they were fossils derived from the Chalk, thus dating the Sponge Gravels as late Cretaceous. Sharpe stated his conclusions in a paper read to the Geological Society in 1854 and deposited his specimens with them. His friend Davey said that most of them had disappeared by 1874, but the types of a new species, *Idmonea hagenowi*, described in the paper have survived and are now in the

Geological Survey collection. Sharpe was born at Marylebone in 1806, orphaned in infancy and brought up by a half-sister. He entered the office of a Portuguese merchant in London at the age of sixteen. His interest in geology commenced when he was twenty-one and he joined the Geological Society. After a visit to Portugal he left the Portuguese merchant and joined his brother, also a merchant in London. Later visits to Portugal resulted in papers on the geology of the country around Lisbon and Oporto. He also did some good stratigraphical work on the British Palaeozoic, some rather controversial work on the Alps, and wrote several papers on palaeontology, but only the one dealing with Polyzoa. He died a bachelor in 1856.

At this point it would be useful to mention his friend, the Rev. Henry Davey, although he was a younger man and rightly belongs to the latter half of the century. He duplicated Sharpe's collecting at Faringdon and published privately a fossil list, in time, he tells us, for the visit of the Geologists' Association to Wantage on 7 May 1877. He asked the French palaeontologist de Fromentel to name his specimens, and accepted Neocomian or Aptian age for the Sponge Gravels. Unfortunately for Davey, but fortunately for us, he got into financial difficulties through losing a lawsuit and had to sell his collection. It was bought by the University Museum at Oxford, where it now is.

While the marine and fossil Polyzoa were both receiving attention there was a worker on the freshwater species who made fundamental contributions to the classification of the Polyzoa. George James Allman was born at Cork in 1812 and intended to take up law. Fortunately he changed his mind and graduated in Arts and Medicine in Dublin at the late age of thirty-six. He was successively Professor of Botany at Dublin and Regius Professor of Natural History at Edinburgh, resigning owing to ill-health in 1870. That however did not prevent him from undertaking the Presidency of the Linnean Society from 1874 to 1881 and the British Association in 1879. He divided the Ectoproct Polyzoa into the Phylactolaemata, where the tentacles are arranged in a horseshoe, and the Gymnolaemata, where they form a complete circular crown, the first being confined to freshwater, the second mainly marine but with a few freshwater species.

The richest fauna of fossil Polyzoa in this country is of course in the Coralline Crag and the early work on it was by Searles Valentine Wood, Senr. He was so named because he was born on St. Valentine's Day in 1798 and seemed to like his name so much that he complicated things for posterity by naming his son Searles Valentine Wood also! Wood started his career as a midshipman but gave up the sea because he was disappointed at not receiving a promised command. He decided to devote his life to natural history, settled down in his native Suffolk, and took up palaeontological work, making extensive collections of Crag material which ultimately passed to the Natural History Museum. He was a friend of F. W. Harmer, the Norwich textile manufacturer, and worked with him on problems of East Anglian geology. Later in life poor health compelled him to come to London, where he met and worked with Lyell. At the time of his death in 1880 he was Treasurer of the Palaeontographical Society and was succeeded in that office by his son. The son followed the father in other ways; he gave up his career as a solicitor in Woodbridge for scientific work and, like his father, worked with Harmer in East Anglia.

Intermediate to the two S. V. Woods was George Busk, born in 1807 at St. Petersburg, who was to use the collection made by S. V. Wood, Senr. He trained as a surgeon at the Royal College of Surgeons, St. Thomas's and St. Bartholomew's Hospitals, and was appointed to the seamen's hospital ship *Dreadnought*, moored in the Thames Estuary. Possibly his choice may have been due to a desire to see the world, but he does not appear to have even had the chance to see the sea, and after twenty-five years he resigned in 1856 to devote himself to scientific work. He had already written a monograph on Recent Polyzoa in 1852 and an article on the subject in the *English Cyclopædia* in 1856. He was asked to undertake a monograph of the Crag Polyzoa, a work which Haime was to have done but for his early death. Busk tells us that he started it with some misgivings at following so eminent a palaeontologist, but the monograph of the Crag Polyzoa, published in 1859, was a very good piece of work and it is still the only major work devoted entirely to the Crag Polyzoa. It is essential to all work on the late Tertiary and Pleistocene. Much of the material Busk used was from the collection made by S. V. Wood. Busk was an artist and his figures superb; he was the first worker on the Polyzoa to draw realistic representations of his specimens. Other earlier workers gave stylised figures which possibly showed the points they wished to emphasise but did not represent the actual specimen. I have had the pleasure of seeing some of Busk's sketches of Recent Polyzoa, all made in pencil on small pieces of paper and in all stages from rough drafts to finished drawings. The obvious care lavished on these sketches explains the excellence of his published figures. Busk also made hectograph copies of his figures which he sent to other workers and many of these are still in existence. Busk's interests were wide, and he made a success of the many things he attempted. He was elected a Fellow of the Royal College of Surgeons in 1843 and became President in 1871. He was at one time editor of the *Microscopical Journal* and later in his life was elected President of the Anthropological Society. He was also Treasurer of the Royal Institution. When experiments on animals came under Government control he was one of the first inspectors appointed. His last major work was the *Report on the Polyzoa of the Voyage of the 'Challenger'*. This he completed while contending with failing health and he was just able to read the proofs before he died in 1886. He was too ill in 1885 to receive personally the award of the Wollaston Medal.

In the latter half of the nineteenth century many workers were engaged on the description of fossil Polyzoa. de Loriol published on the Neocomian and Valangian fauna of Switzerland, Beissel on the Maastrichian, Ottomar Novák on the Cretaceous of Bohemia, Pergens and Meunier on the Cretaceous of Belgium and Hennig on the Swedish Cretaceous, A. Manzoni dealt with Italian fossil Polyzoa and the Miocene species of Austria and Hungary. In the United States William More Gabb was also working on Cretaceous and Tertiary faunas in California when a field geologist's life was reminiscent of a Western film, his party riding into wild country on horseback with everything they needed on their saddles and frequently changing their route to avoid hostile Indians. Gabb was born in 1839 and received a classical education. He had an interest in minerals and decided to seek out the great American geologist James Hall and ask his advice on a career. For a time he was Hall's assistant and pupil but in 1862 he was selected to serve on the newly formed California Survey on account of his knowledge of Cretaceous palaeontology. After the publication of his work on Californian fossils, including the Polyzoa, partly in collaboration with Horn he left the State to become geologist to a mining company in San Domingo in 1869. In 1873 he was appointed Geologist to the Costa Rican Government but work in a malaria infested country weakened his health and he died in 1878 without having had the chance he wanted to publish completely his palaeontological work.

James Hall, geologist to the fourth district of New York State, described the rich fossil faunas of the Devonian in many papers and monographs—he published around 250 including many on Polyzoa. In his work he was assisted among others by his nephew, George B. Simpson. Hall was born in 1811 and was fortunate in being educated at one of the few schools then teaching Geology in the States. He served the one State for almost the whole of his career and to him fell the honour of being elected the first President of the Geological Society of America. He died in 1898.

While Hall was working on the North American Palaeozoic, H. Alleyne Nicholson was describing Corals, Sponges and Polyzoa from the State of Ohio. He was Professor of Natural History at Toronto from 1871 to 1874 and his work was published in 1875 when he left for St. Andrews. There he continued his work on Palaeozoic fossils, publishing his monograph *On the Structure and Affinities of the Genus Monticulipora* in 1881. Subsequently he was Regius Professor of Natural History at Aberdeen.

One interesting piece of work was the description by W. Keeping of Cambridge in 1883 of the Polyzoa from the Neocomian of Brickhill and Upware. The deposit has now been completely quarried away, and but for his collection in the Sedgwick Museum there would be little trace of this fauna today.

So far, all the work on Polyzoa had been done by men, but now the first woman comes into the story. Miss Eliza Catherine Jelly was a link between workers in many parts of the world—she corresponded widely and was a channel for the exchange of material. Her *Synonymic Catalogue of the Recent Marine Bryozoa* was published in 1889, and included fossil synonyms. It is remarkable for its completeness and accuracy. She had at some time worked over Davey's collection of fossils, though whether that was when he had it or after it went to Oxford is doubtful. She gave a collection of over 1100 specimens to the Manchester Museum and in 1896 she gave the residue of her collection, some 2000 mounted specimens of Recent Polyzoa, to the University Museum at Cambridge when she had had to give up her work on Polyzoa owing to age and eye trouble. In the following year she sent her manuscript notebook to Sir Sidney Harmer, telling him to keep it if it was of any use to him. He did keep it and it is now in the Harmer Library at the Natural History Museum. A note inside the cover explains that where she was unable to obtain papers on new species for herself she borrowed them, traced the figures and copied out the descriptions. In other cases the authors had helped her—Busk sent many of his hectographed figures and Waters, of whom we shall hear later, sent photographs. Her tracings were most carefully done and tell something of the painstaking worker she was. Other than her one published work and a few letters we know nothing of her; she faded from Polyzoan scene and no one wrote even an obituary on her. However, Pergens, Harmer and D. A. Brown have all named species in her honour.

Two amateur zoologists who did much to forward the study of both Polyzoa and Hydrozoa were the Rev. Thomas Hincks and Canon Norman. They were both born in Exeter, Hincks on 15 July 1818 and Norman in 1831. Hincks was the son of a Professor of Natural History and probably owed his interest in Polyzoa to an early friendship with Allman. He went into the Unitarian ministry in 1839 and after a succession of churches was appointed to Mill Hill Chapel, Leeds, where the great Joseph Priestley had held office in earlier years. His ministry was very successful, for fourteen years, until his voice failed in 1869. This was a tragedy for a minister, and when rest and treatment both failed to restore it he had to resign. From that time on he wrote a series of papers on Polyzoa from all parts of the world, mostly in the *Annals and Magazine of Natural History*. His first work was a two volume *History of the British Marine Polyzoa*, published in 1880. The many papers on overseas Polyzoa were ultimately gathered up in one volume, *Contributions towards a General History of the Marine Polyzoa, 1880–1891*. He retired to Taunton and later to Clifton, where he died on 25 January 1899.

Alfred Merle Norman, like Hincks, was a minister of religion, but before entering the ministry he held one post which altered the interests of his life.

He was educated at Winchester and Christ Church, Oxford, and then took a post as private tutor with the Dowager Duchess of Glasgow's household at Cumbrae from 1854 to 1855. There he became interested in marine fauna and spent his summer vacations dredging. After he entered the Anglican ministry he still retained the interest he had acquired in Polyzoa and Hydrozoa. He held a succession of livings, until, like Hincks, he had to retire owing to ill-health. He was made an Honorary Canon of Durham. He was, however, able to continue his scientific work, published many papers, and in spite of poor health he lived to the very reasonable age of eighty-seven, dying on 26 October 1918.

Strangely enough, work on Polyzoa seems to have thrived on ill-health. Contemporary in part with the last two zoologists was Arthur William Waters, who was born at Alderley Edge about the middle of last century and was intended for the family business. To that end he studied Engineering and Chemistry at Manchester, but unfortunately contracted lung trouble and was sent to Switzerland for the high altitude cure. He lived there for many years and as his health improved travelled on the Continent. His general interest in natural history led to a special interest in fossil Polyzoa. Later he spent some time on the Italian coast and studied Recent forms. He received collections of both fossil and Recent material from Australasia and published papers on both. He also dealt with the Polyzoa collected by several expeditions. In spite of his early troubles he lived to the age of eighty-two, dying on 20 February 1929, and he had used his time to engage in much useful work. Probably on account of his early collecting on the Continent he preferred the name Bryozoa to Polyzoa. He left all his papers and specimens to Manchester University.

Waters's progress from fossil to living material was paralleled by Georg M. Levinsen, a Dane, who was born about the same time, 23 January 1850, and commenced his studies on fossil Polyzoa. He decided that he had first to study living material in order to understand the fossils which interested him. In the course of his work he made useful contributions to the classification and detailed anatomy of the Cheilostomata. He published a number of works on Polyzoa during the latter part of last century and up to his tragic death on 8 August 1914. His monograph on the Danish Cretaceous Polyzoa was published in 1925.

A number of amateur geologists deserve a place in this story of the Polyzoa, some because of their published papers and others for their carefully made collections. George R. Vine, a Fellow of the Geological Society and a member of the Yorkshire Geological Society, published a number of papers in the *Reports of the British Association* between 1880 and 1893. Vine introduced the form Cryptostomata for one order of the Palaeozoic Polyzoa. A. W. Rowe published his work on the Zones of the White Chalk of the English Coast in our PROCEEDINGS in the early part of this century. He was a careful collector and recorder of Chalk Polyzoa and worked in association with another amateur, Frederick Gamble, a prison officer at Maidstone, who collected largely in the Charing and Chatham areas. Both contributed material to the Natural History Museum. Rowe was the first to suggest the use of a dental drill to develop fossils but he and Gamble used dripping water to reveal the fine detail of Polyzoa, a slow method but especially good with Chalk material. Gamble wrote to Rowe in 1911 and concluded his letter with this remark: 'The material I sent you has not been treated with the brush, I consider to do so is injurious to the delicate and important parts such as the delicate ridges of cells and avicularia and pores. Nothing but water from the tap is used by me.' Rowe had been interested in Geology since boyhood and complained that his busy practice as a doctor limited the time he could spend in the field. However, he succeeded in publishing his work before he retired in 1910!

Another amateur worker on Chalk Polyzoa was a solicitor, R. M. Brydone, who wrote more papers on Chalk Polyzoa than any other English worker. When a boy at Winchester he was introduced to the Hampshire Chalk but much of his later work was done in East Anglia. He commenced publication of a series of papers in the *Geological Magazine*, but a serious illness interrupted the series and he published his later work privately. He died at the age of seventy at Mundesley on 6 August 1943.

Llewellyn Treacher was a great collector of Polyzoa though probably better known to most for his flint implements. He left a fine collection of Faringdon material to Reading University. He could not acquire some works on Polyzoa which he wanted but was able to borrow them. He made some surprisingly good photographs of the plates with a pin-hole camera but apparently did not make copies of the descriptions, so that it is difficult to see how he used them. These photos passed from him to the late A. G. Davis and I now have them as a memento of a great amateur and a reminder of the difficulties which many like him have faced.

Arthur G. Davis was known to many of us for his work on the Lower Tertiaries and their faunas. It was his intention to monograph the Tertiary Polyzoa when he retired but he died before the chance came. Through his interest in the Polyzoa he discovered the first fossil *Rhabdopleura* in the London Clay of Lower Swannick, Hampshire. Allman had placed Recent species of *Rhabdopleura* in the Polyzoa but later workers had indentified the genus as a Pterobranch. Davis had made an important discovery and the species was figured and described by Thomas & Davis in 1949 and named *Rhabdopleura eocenica*. Davis also published a paper on the English Lutetian Polyzoa in our PROCEEDINGS.

P. H. MacGillivray carried out much work on Australian Polyzoa, contributing on Recent species to the *Natural History of Victoria*, published under the editorship of McCoy, and writing a monograph of the Tertiary

Polyzoa of Victoria which was published just after his death in 1895.

The name of J. W. Gregory is known in many fields of geological work. His first appointment was at the Natural History Museum, where he dealt among other things with the fossil Polyzoa. John Walter Gregory was born in Bermondsey on 27 January 1864, the son of a wool merchant. He was intended for a commercial career but felt a strong urge to work on natural history. Fortunately he followed it. He produced the *British Museum Catalogue of the Jurassic Bryzoa* in 1896, and two volumes of the *Cretaceous Bryzoa* in 1899 and 1909. These are standard works for the Cyclostomes and include descriptions of many Continental species as well as British ones. In the volume on the Jurassic he discusses the problems of classification of tubular fossils and Cyclostomes in particular. He was critical of the artificial divisions which earlier workers had used and made useful contributions to the classification of Cyclostomes.

Later in life he travelled widely on geological work and wrote some 300 papers and twenty books, including some of a popular nature. He met an untimely death by drowning while on geological exploration in Peru on 2 June 1932.

In the field of zoology very important work was carried out by Sir Sidney Harmer, born on 3 March 1862, son of the F. W. Harmer we have already mentioned as a co-worker with the Searles V. Woods. F. W. Harmer was a textile manufacturer and wool merchant but very wisely did not press his son to follow in the business. The son later acknowledged his ather's encouragement in natural history work. He had a distinguished areer at Cambridge and at the Natural History Museum. He was President of the Linnean Society from 1927 to 1931. His work on the Polyzoa colected by the *Siboga* Expedition was his great contribution to the subject nd is a standard work now. In addition to his systematic work he disovered embryonic fission in the Cyclostomes. He died on 22 October 1950.

Contemporary with Harmer, Folke Borg was investigating the Recent Cyclostomes in Sweden and made careful studies of the mode of growth of he zoarium. He also worked on the problems of the Heteroporidae. Most f his work is published in English and is a very good introduction to the study of fossil Cyclostomes.

So far I have only dealt with workers who are no longer with us, but it would be a very incomplete account of the work on Polyzoa which did not mention one living American and some of his colleagues. Dr. Ray S. Bassler, either alone or in collaboration with others, has worked on both ossil and Recent species (sometimes on both in the same monograph). When a schoolboy in Cincinnati he went out with the Cincinnati Society of Natural History and collected Palaeozoic Polyzoa from the richly fossiliferous Ordovician of that area. This brought him into contact with Ulrich, he author of many papers on North American Palaeozoic species. Ulrich earned his living by selling fossils and while still at school Bassler joined him as a part-time assistant. The Society allowed him to browse in their library, where he found Miss Jelly's *Synonymic Catalogue of the Recent Marine Bryozoa*, and was fired with the idea of doing the same for the American fossil species. Later in life he did this for the Palaeozoic Polyzoa in collaboration with J. M. Nickles. He is also the author of Volume G (Bryozoa) of the recently published *Treatise on Invertebrate Paleontology*. His first official appointment was at the Smithsonian Institution where he worked on Polyzoa, and later he and Ferdinand Canu of France were entrusted with the formidable task of working out the North American Tertiary Polyzoa. This and much other work was completed before they first met each other in 1926. All their communication had been by post. Even during the First World War period they still continued to send material and manuscripts backwards and forwards across the Atlantic, and none was lost. Their partnership was only ended by Canu's death from cerebral haemorrhage in 1932. Canu worked to the end, completing his part of a joint paper on the Tertiary Cheilostomes of Victoria and posting it back the day before he died. It was an outstanding alliance in both the fossil and Recent fields and their output of work on fossil Polyzoa is greater than that of anyone else. Their one monograph on British species was on the fauna from the Sponge Gravels of Faringdon.

Those of us who were on the Austrian Field Meeting will remember Dr. Bobies who led us on the Miocene at Eisenstadt and St. Margareten. He carried out the most recent work on the Austrian Miocene Polyzoa and the final part of it was only published after his death. Probably his day with us was one of the last occasions he had in the field.

This has been a brief picture of some of the members of the team in time who have worked on the fascinating subject of Polyzoa. Some made the study their vocation, some were so drawn to it that they threw up their professions to pursue it, some struggled against ill-health to carry on their work, and some took it up because ill-health had closed other outlets for their energies. All had the desire to understand living things whether of the past or the present.

ACKNOWLEDGMENTS

I should like to thank Dr. A. B. Hastings and Dr. H. Dighton Thomas for making unpublished material available and Dr. Goodhart of the Museum of Zoology, Cambridge, for information regarding Miss Jelly.

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Contributions to the Geology of the Mere Fault and the Vale of Wardour Anticline

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Revised version received 20 April 1959

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ABSTRACT: Detailed mapping of the western end of the northern limb of the asymmetrical Tertiary Wardour Anticline around East Knoyle and the Mere Fault from West Knoyle to Bayford, a distance of nine miles, is described. The Mere Fault downthrows north bringing Jurassic rocks on the upthrown side against Cretaceous Beds on the downthrown side. The fault throw ranges from 600 feet at West Knoyle to 400 feet at Mere and 150 feet at Penselwood and Bayford. It maps as a steep reversed fault and as it is closely associated with the Tertiary Wardour fold it is likely to be of similar date. In this respect the Mere Fault has close analogues with the Poxwell–Ridgeway faults in south Dorset and with the Hog's Back structures in Surrey.

1. INTRODUCTION

THE VALE OF WARDOUR is where a typical asymmetrical Tertiary fold, rending nearly east–west, has bent upper Mesozoic strata upwards so that subsequent erosion has stripped off the Chalk cover to expose all formaions beneath from the Upper Greensand to Kimmeridge Clay at the surface see Fig. 1). The Upper Greensand, Purbeck and Portland Beds, which outcrop within the Vale of Wardour, are exposed in many quarries from which stone was taken in an area where good building material is scarce. The Mere Fault is a post-Cretaceous reversed fault, downthrowing north and extends generally westwards from the Vale of Wardour near East Knoyle to Mere and Wincanton. Chalk and Albian, on the downthrown ide, are brought against Kimmeridge Clay and Corallian on the upthrown ide. The fault and fold appear to be two different surface expressions of

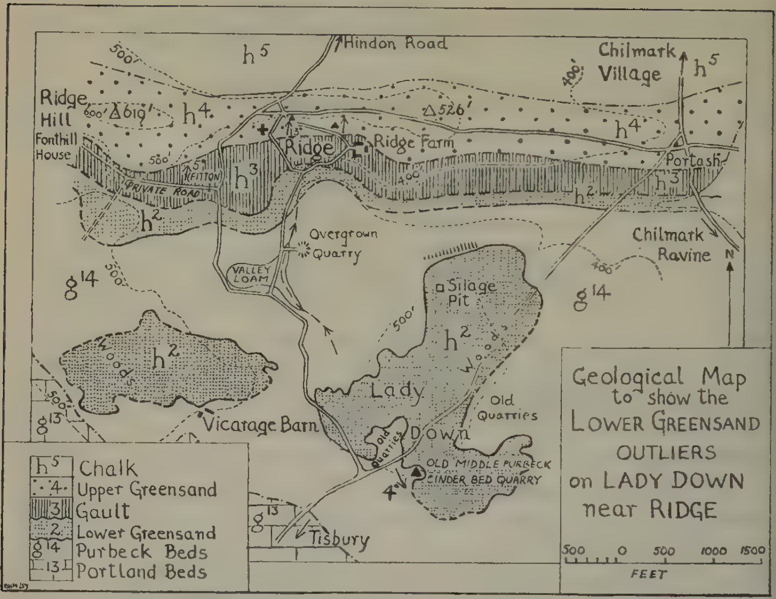


Fig. 1. The boundaries of the Lower Greensand outliers on Lady Down were mapped by the author in May 1957. The other boundaries, east and west of Ridge, are based on Reid's mapping in 1900, but modified, where continuous lines are shown, by the author

yielding to a general pressure from the south in this part of south-west Wiltshire and north Dorset during the Tertiary folding.

(a) Table of Formations and Useful Exposures

Below are given some of the more useful exposures, some occurring outside the map showing the Mere Fault (Fig. 2).

Formation	Thickness in feet	Useful Exposures
Upper Chalk	800+	West Harnham, Bemerton, Charnage, East Knoyle.
Middle Chalk	80–100	White Sheet Hill (both), Fovant, Buxbury Hill.
Lower Chalk	180	
Warminster Greensand and related beds	6–10	Mere, Upton, North Melbury, Gains Cross (in the Stour Valley, north-west of Blandford) (846107).
Upper Greensand	150	Baverstock, Dinton, Fovant, Shaftesbury, Upton, Wolverton.

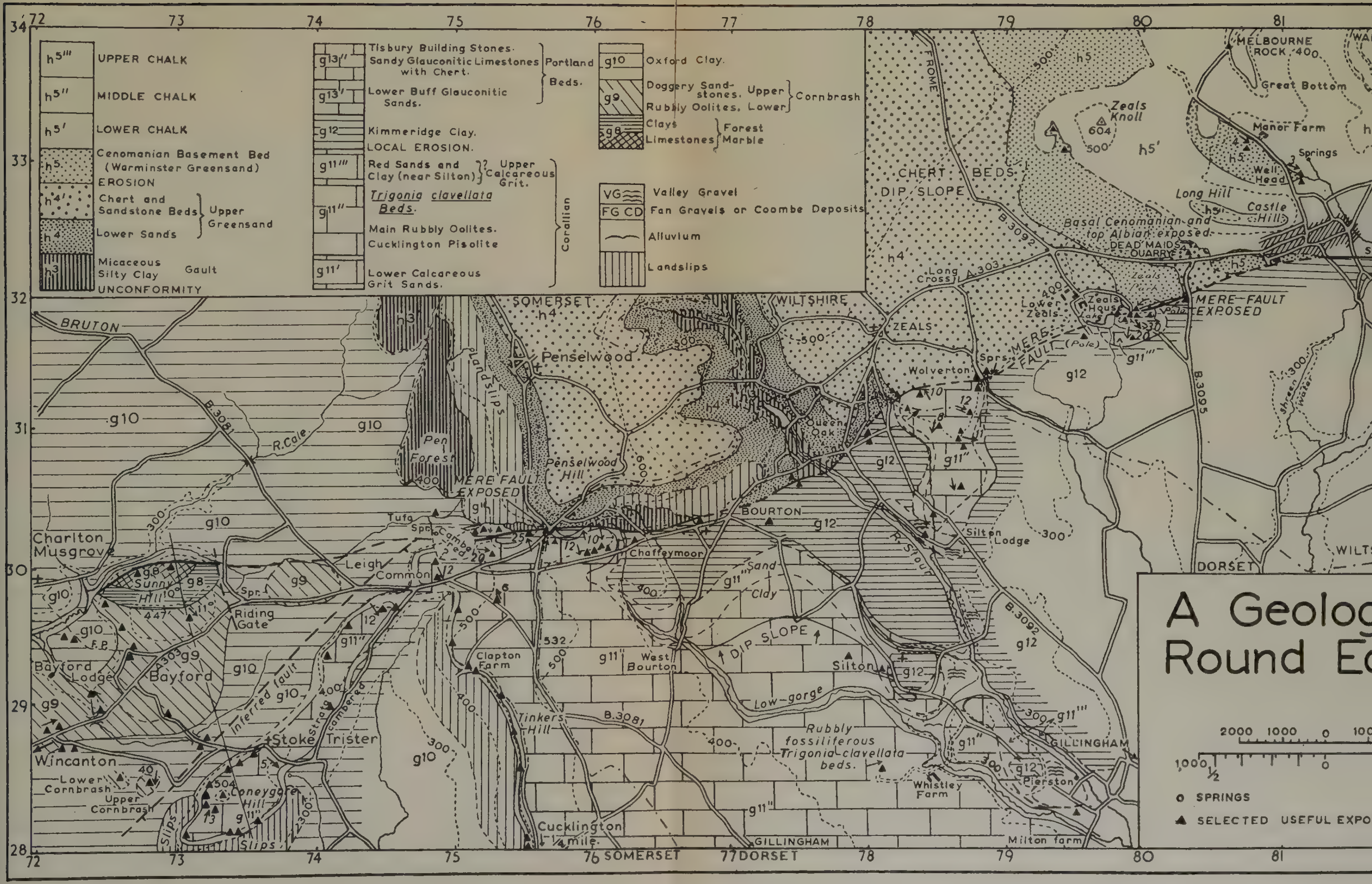
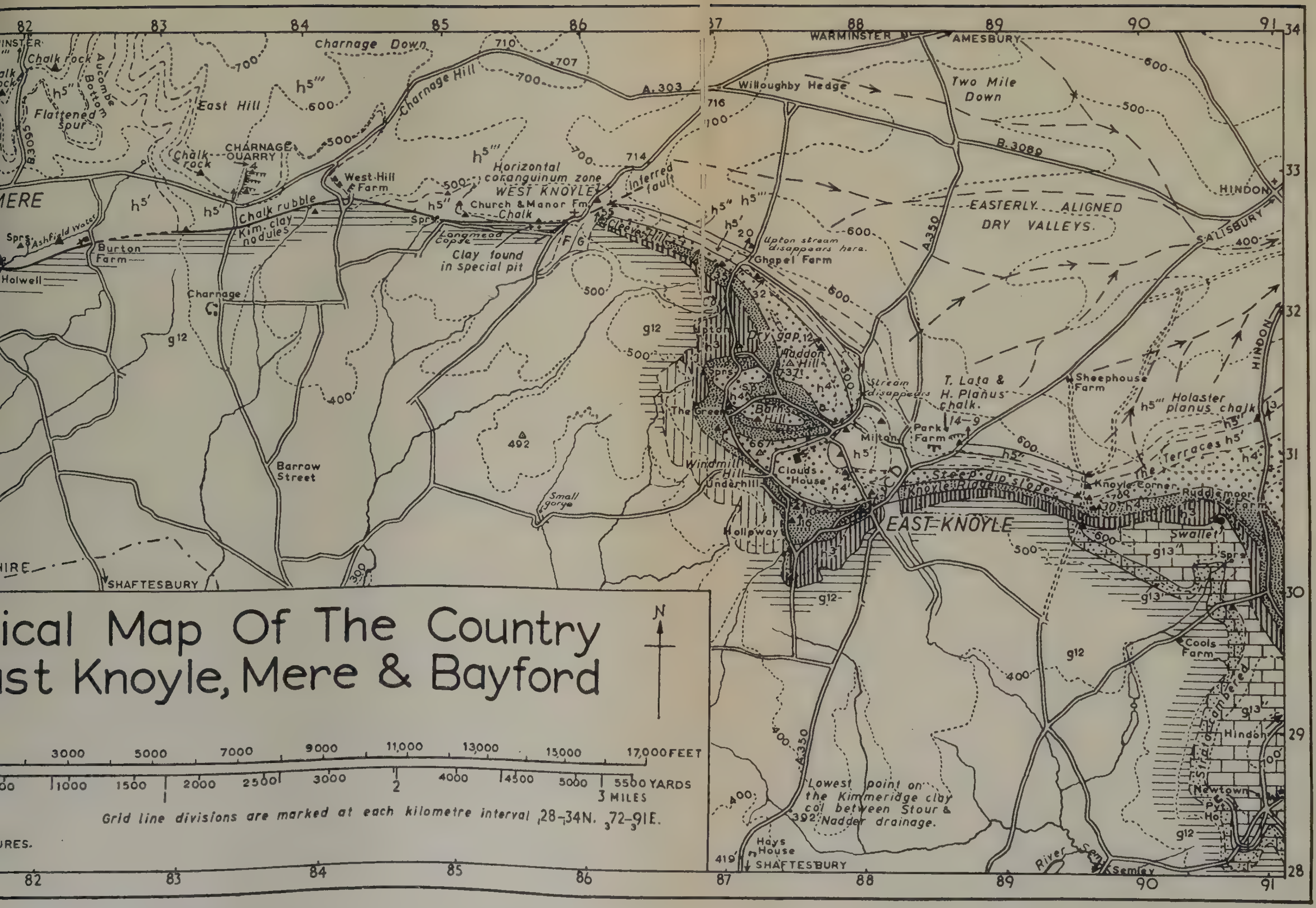


Fig. 2. The map and notes are based on a six-inch survey by the author in 1951–2 and 1954 with these additions: Michael House as
of the Mere Fault and Chalk between Charnage Quarry and Mere are based on F. H. Edmunds (1938). East of a north–south line t



sted in mapping the Corallian near Clapton Farm and the small Cornbrash outcrops south of Bayford in 1954. Boundaries
ough Knoyle Corner the boundaries are based on C. Reid (1900) six-inch maps with small alterations

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Formation	Thickness in feet	Useful Exposures
Gault	75–90	Dinton Brick Yard, Fovant Cress Beds, Wardour Park (basal beds).
Lower Greensand	0–25	Wardour Park (and surface rubble south of Tisbury) and between Ridge and Lady Down.
Wealden	0–40	Formerly west of Dinton Station.
Upper Purbeck	22	
Middle Purbeck	23	Teffont, Chicksgrove, Lady Down.
Lower Purbeck	65	Chicksgrove, Teffont, Oakley, Chilmark.
Portland Beds	100	Chilmark, Fonthill Park, The Lawns, Oakley, Court Street, Newtown, Tisbury.
Kimmeridge Clay	650(?)	Gillingham Brick Pit (Pictonia Zone), also in cutting slips near Semley Station.
Corallian Beds	100–120	Marnhull, Todber, West of Gillingham, Cucklington, along the Mere Fault.
Oxford Clay	600(?)	(At Westbury, Wilts, boring.)
Cornbrash	25	Wincanton and Bayford.
Forest Marble	top 100	North side of Sunny Hill, Bayford.

Generally the highest Jurassic rocks are only exposed near the eastern end of the Vale, as the Aptian and Albian transgress westwards on to older Jurassic strata. Thus at Dinton, basal Wealden is preserved beneath the local Aptian, but by Tisbury Gault rests on Portland Beds, on Kimmeridge Clay around Shaftesbury and East Knoyle and by Penselwood Hill, five miles west of Mere, Albian rests on Oxford Clay.

(b) Brief History of Previous Work

The great variety of formations exposed within the Vale of Wardour has attracted geologists since early in the nineteenth century and there are many descriptions by many writers.

Lady Bennett provided Sowerby (1818) with some of the earliest ammonites described from the Portland Beds of Chicksgrove (south) quarry, and also referred to the Tisbury Star Coral (*Isastraea oblonga*). The first comprehensive account was that of Fitton (1836). He noted the sandy nature of the Chilmark building stones, found Purbeck dirt beds and discovered a cycad trunk near Tisbury. He observed the Hastings Sands (Wealden) above the Purbeck Beds near Dinton and separated a sandy bed below the Gault which was later assigned to the Lower Greensand. Fitton also realised that the Wardour fold was asymmetric, with steeper northern dips, and included a section diagram with his account. In 1856 there appeared the first one-inch Geological Survey maps of the area which had been surveyed by Bristow.

In 1877 Blake & Hudleston gave the first comprehensive account of the Corallian outcrop in north Dorset and were able to link up the north Dorset succession with that of the type locality at Weymouth. They described the apparent northward thinning of the Upper Corallian. They also noted the increasing number of rolled corals in the upper beds in the same direction, and commented on the false-bedded Todber Freestones. Three years later this was followed by a study of the Portland Beds within the Vale of Wardour. In this paper they established, amongst other facts, that the Upper Freestone Building Stones fifteen feet thick in the Chilmark Ravine are reduced to a two-foot band, crammed with *Camptochlamys lamellosus* at Chicksgrove and Oakley, within only one and a half miles. The nature of the junction with the overlying Purbeck has been much discussed since then, and is still not settled. In 1881 W. H. Hudleston led the first Geologists' Association excursion to the Vale of Wardour and in the same year the Reverend W. R. Andrews (1881), then resident at Teffont, published the first account of the presence of the Middle Purbeck marine Cinder Bed in the Vale of Wardour. In 1894 this was followed up by a comprehensive description of the whole Purbeck sequence (Andrews & Jukes-Browne, 1894), based on the Dorset ostracod divisions, but this was disregarded by Woodward (1895) when writing his Survey Memoir.

By 1900 the Geological Survey completed the six-inch mapping for the New Series one-inch map, Sheet 298, which includes that portion of the Vale of Wardour east of Tisbury. In the accompanying memoir (Reid, 1903) he firmly established the presence of both Wealden and Lower Greensand between the Gault and Jurassic beds. He also records seeing the Tisbury Star Coral *Isastraea oblonga*, already recorded by earlier writers, but in position of growth. This was observed in what must have been a temporary section, where a lane forks off the road from Tisbury to Fonthill, three-quarters of a mile north-west of Tisbury Square.[1] On the other hand, Reid adhered to Woodward's interpretation of the Purbeck succession and discounted the unconformable Wealden boundary suggested by Andrews & Jukes-Browne. The latter writer (Jukes-Browne, 1903) added further material on the Purbeck–Wealden boundary and it appears that Reid's mapping followed Woodward's Purbeck divisions, and needs some revision to fit in with the palaeontological divisions now used. In the same year (1903) as the memoir appeared, there was a second Geologists' Association excursion to the Vale of Wardour (Blackmore & Andrews, 1903), and by 1904 Jukes-Browne had completed his survey of the Cretaceous rocks (1900–4), which includes descriptions of sections west of the area dealt with in the sheet memoir, some of which do not appear to have received any attention since. Jukes-Browne included a quarter-inch map, in his *Cretaceous Rocks*, Part I (1900), showing the Mere Fault, but no detailed mapping had been done.

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[1] Information (*in litt.*) from Mr. F. H. Edmunds in 1951.

In the fifty years since 1904 there have been only a few further references to this interesting area. In 1933 Dr. Arkell gave an admirable summary of the Jurassic rocks; he followed up the earlier observations on the dissimilarities between the Chilmark–Tisbury building stones and the Dorset counterparts, and attempted to disentangle the confused ammonite nomenclature of the Portland Beds. He placed, tentatively, but almost certainly rightly, the main Tisbury and Lower Chilmark building stones in the upper part of the Portland Sands of the Dorset coast, only retaining the oolitic upper Chilmark building stones within the Dorset Portland Stone. This will entail the remapping of the Wardour Portlandian to fit into the new classification. (See also Arkell, 1935, for correlation table.) (House, 1958.) In 1938 F. H. Edmunds added a contribution to the physiographical evolution of this area which accompanied the fourth Geologists' Association's Field Meeting to the area. The next year, J. F. Kirkaldy (1939) refers to the thin Lower Greensand below the Gault that crops out round the Vale, and also south of Shaftesbury, but he was unable to determine the zonal position of either outcrop.

Outside the Vale of Wardour proper, the Warminster Greensand beds at the base of the Chalk Marl have received attention from Jukes-Browne in 1896, 1900–4 and 1901, and from Scanes, jointly with Jukes-Browne, in 1901, and with Pope-Bartlett in 1916 when, in the latter year, both authors led the third Geologists' Association excursion. All the earlier Warminster Greensand accounts have been fully summarised by Edmunds (1938) for the fourth Field Meeting in the area of the Geologists' Association in 1937. It now seems clear that the fossils from the Warminster Greensand are Cenomanian in age and the majority did not come from Warminster itself but from Maiden Bradley and Mere. Nevertheless, the Warminster name has been adopted for these beds, whereas today the best available section is at Dead Maid Quarry, Mere. Also included in Edmunds' account (1938) is the first contribution on the Mere Fault which was the present author's starting point for detailed mapping. Edmunds estimated the Mere Fault to have a northerly downthrow of about 600 feet near Charnage Quarry. He shows the fault to be reversed, passing just to the south of Mere, and downthrowing Lower and Middle Chalk against Kimmeridge Clay.

Wooldridge & Linton (1955) have given the whole area prominent attention in their comprehensive survey of the *Structure, Surface and Drainage of South-East England*. They regard the three east–west lines of downland ridges comprising (a) The Great Ridge–Mere White Sheet Hill ridge, and (b) The Barford Down–Berwick St. John White Sheet Hill ridge, and (c) the Melbury Beacon, Win Green–Coombe Bissett Down ridge as type examples of the remnants of the Mid-late Tertiary Peneplains. Also they regard the Wardour drainage as now adjusted to structure through two cycles of erosion and that the bulk of this area lay outside the furthest advances of the Pliocene Sea.

At Whitsun in 1954 this writer led the fifth Geologists' Association Field Meeting over the Vale of Wardour and the Mere Fault country from Shaftesbury. The Field Meeting report (Mottram *et al.*, 1957) included some of the more interesting localities visited from which the palaeontological records were obtained by various people, as well as brief references to the Mere Fault at West Knoyle, Charnage, Mere, Wolverton and Penselwood Hill, visited by the Association during the excursion.

(c) Tectonic Summary of the Wardour Anticline

The tectonics of the Wardour fold have not been described in detail previously, but much is self-evident from the New Series one-inch map, Sheet 298, which shows the Vale of Wardour as far west as Tisbury and Ridge. The Wardour Anticline has an amplitude of about 1200 feet so that around Tisbury the Cenomanian base must have risen to about 1000 feet above present O.D. The fold has steeper northerly dips than those on the southern limb, which are everywhere from 3–5° south (see Fig. 3, Section 1). The northerly dips gradually steepen westwards and west of the Fonthills a 10–15° NNW. dip can be seen in *planus* Zone chalk on the roadside from Tisbury to Hindon. The accompanying map to this paper (Fig. 3) shows how the northern limb continues to steepen westwards past East Knoyle to where the Mere Fault begins at West Knoyle. North-west of East Knoyle, around Windmill Hill, the Green and Upton, there is quite clearly a local roll displayed by the Upper Greensand outcrops. These double back eastwards from Windmill Hill to Milton before resuming their north-west trend along Haddon Hill. Also the Gault base rises above the 600-foot contour west of Upton, so in this small and interesting upland area north-west of Clouds House there is a periclinal fold, pitching east, riding on the main northern limb of the Wardour fold (see Fig. 3, Section 2).

In general the Jurassic rocks are nearly conformable to the overlying Cretaceous, but these are local variations in addition to the steady westward Cretaceous overstep. The Jurassic rocks within the Vale of Wardour are affected by a series of gentle rolls, trending north-west to south-east, which disappear under the transgressive Lower Greensand and Gault without affecting them. The Portland Beds, exposed in the Chilmark Ravine, are brought up by a shallow anticline and this is flanked to the south-west by a shallow syncline which brings the Wealden down to river level again near Sutton Mandeville. The next undoubted flexure affects the Portland Beds south-east of Knoyle Corner, but is only partly preserved beneath the transgressive Gault above, and appears on the map (Fig. 2).

Around Tisbury itself the numerous Portlandian quarries show a variety of dips. Some apparently can be ascribed to false bedding. This was seen in 1940 and 1941 in the temporary opening of a shallow quarry between road and railway 500 yards north-east of Hazeldon Farm (936381). False bedding was also seen in 1949 in a new track section south of Court Street. However, in other pits, the dips appear to point towards the valleys. Some of these exposures show considerable gulling, like those that can be seen in Tisbury West Quarry on the Newtown Road, and still being worked. This gulling recalls the cambered structures described in the Midland Ironstone field and in the Oxford region by Hollingworth, Taylor & Kellaway (1944) and Arkell (1947a) respectively.

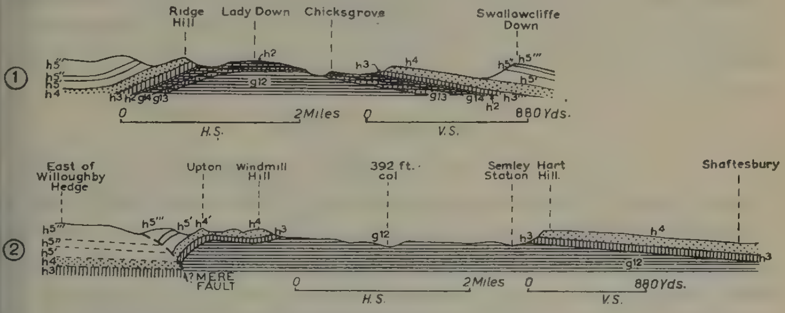

Fig. 3. Sections across the Mere Fault

It appears from notes on Reid's 1900 six-inch maps in the Geological Survey Library, that much, presumably Lower, Greensand debris still remains on the Purbeck outcrop around Lady Down and on the Portland outcrop north-west of Tisbury. This writer was able to map two definite outliers of Lower Greensand on Lady Down and around Vicarage Barn, as shown on Fig. 1.

The Lower Greensand, forming these two outliers, is thin, as the silage pit, four feet deep, reached the base of ferruginous sands, which also contained occasional small quartz pebbles. Lumps of a very dark and hard ferruginous sandstone, recalling a tropical laterite, can also be found with ironstained Purbeck slabs in the surrounding arable fields. This thin veneer-like Lower Greensand outcrop north-east of Tisbury suggests that the present Wardour Jurassic surface round Tisbury may be, in part, an exhumed pre-Lower Greensand erosion surface. This is supported by finding further quartz pebbles and chert debris in arable fields on the Portland outcrop north-east of the cross-roads (at 924293) on the Tisbury to Newtown road. This is south of where Reid mapped the Lower Greensand being overstepped westwards by Gault near the 'Beckford Arms'. It is possible therefore that the existing post-Jurassic material on the Portland Beds is the ultimate remains of the combined residue of Lower Greensand and basal Gault hereabouts. Across the Nadder Valley over the ground north-east of Wardour Castle the Portland dip slope disappears under rounded swells of Lower Greensand and Gault above, before the land rises up towards the Upper Greensand escarpment behind. Walking over this ground the boundaries can be mapped by numerous scattered polished pebbles, many about the size of peas, but some occasionally larger, littering the arable fields. The basal Gault pebble bed is exposed in a small pit in Wardour Park (934277) (see Mottram *et al.*, 1957).

The gentle folding and erosion of the pre-Aptian rocks within the Vale of Wardour probably occurred at the same time as occurred elsewhere in southern Britain, notably that described in south Dorset and in the Oxford area by Arkell (1947 and 1947a). This was in the interval between the deposition of the fresh water Wealden Beds and the first marine Aptian strata.

2. THE MERE FAULT

(a) Description

The map (Fig. 2) shows that at West Knoyle the steepening dips in the Greensand and Chalk in the northern limb of the Wardour fold are suddenly replaced by a large fault downthrowing north. This fault brings flinty Upper Chalk against Kimmeridge Clay. The exact manner of the emergence of this fault from the east is obscured by fan gravels at the mouth of the Chalk Marl strike valley east of the church. It is clear that west of the church all formations from basal Upper Chalk to the Kimmeridge Clay are missing in surface outcrop. The Albian, Gault and Greensand and also the Chalk Marl and Middle Chalk are cut out by faulting so that the Upper Greensand does not reappear till west of Mere, a distance of five miles. This fault was called the Mere Fault by Edmunds (1938). He states that the Mere Fault downthrows north and must have a throw sufficient to cut out all the Upper Cretaceous formations below basal Upper Chalk, a total of 530 feet at least. As an unknown amount of Kimmeridge Clay has to be added to that figure (pre-Gault erosion prevents an accurate assessment) it is probable that the fault throw could exceed 600 feet. Thus the Mere Fault is one of the largest faults affecting Cretaceous rocks in southern England.

West of West Knoyle mapping by this writer shows that the fault trace bends slightly northwards across spurs and southwards across valleys. This shows that the Mere Fault is a steep reversed fault downthrowing north. From West Knoyle to Mere and beyond, exposures, supplemented by augering, enable the fault trace to be fixed within a few feet in several places. The throw diminishes westwards for at the Holwell springs, just east of the town, Lower Chalk Marl is brought against Lower Kimmeridge Clay, suggesting that the fault throw can be up to 400 feet. West of the town downthrown Albian beds rise to the surface at Dead Maid Quarry, well known for its exposure of the Warminster Greensand, Cenomanian in age, at the base of the Lower Chalk. This quarry was begun within 100 yards north of the fault line, to extract Greensand as building stone for the building of much of the older part of Mere.

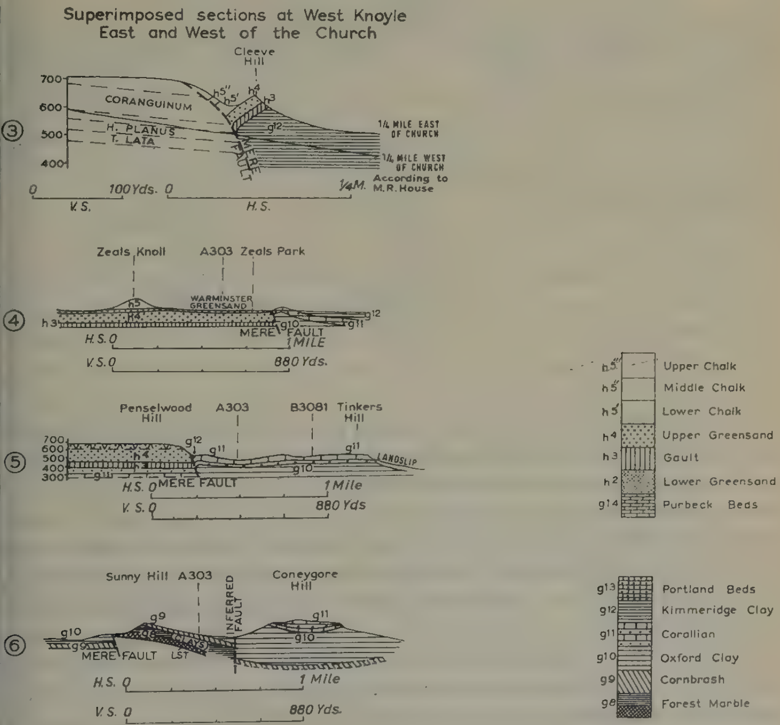

Fig. 4. General sections across the Wardour Fold

In Zeals Park, almost a mile west of Mere Church, there is an isolated hill (31/800318) formed by updomed rubbly oolites of the Corallian *Trigonia clavellata* Beds and cut through by a park drive. Along the northern side of this hill the Corallian oolites are cut off by the Mere Fault and beside the drive the actual fault plane can be seen. The fracture is nearly a clean break, and there is only a little drag occurring within a few feet. The updomed Corallian within Zeals Park, and in the larger inlier, between Wolverton and Queen Oak, south of Zeals village, display

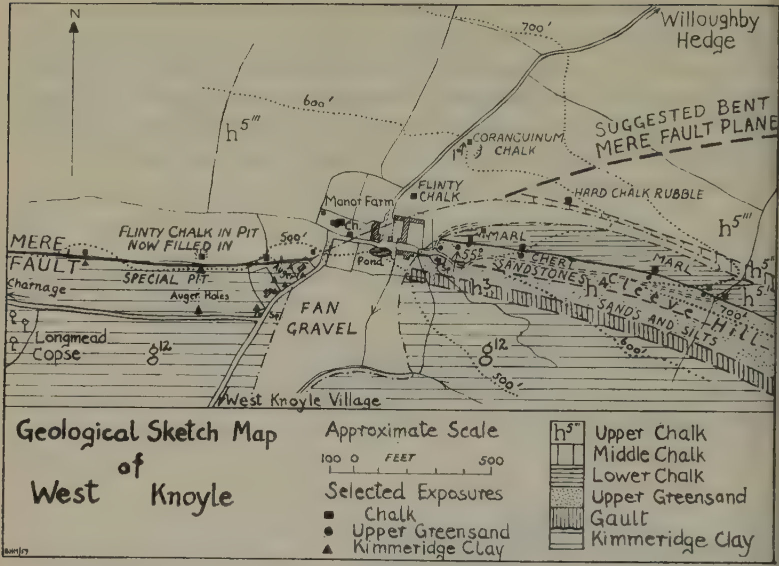

Fig. 5

highly variable dips, suggesting they are local ruckles produced by the overdriven rocks on the upthrown side of the fault (see Fig. 4, Section 4).

The Mere Fault can also be seen at the Penselwood turn (31/757302) on the Dorset–Somerset county boundary. Here Upper Greensand is downfaulted against Corallian, seen, heavily veined with secondary calcite, in an abandoned roadside quarry. Between the two, appearing in the road bank, there is a small sliver of Kimmeridge Clay, pinched in close to the fault, which yielded *Ostrea delta* (Fig. 4, Section 5). North of the fault the Corallian quickly disappears under transgressive Gault for, under Pen Forest, the Gault oversteps on to the Oxford Clay and no oolites of any sort could be found. At Penselwood Hill the fault throw must be about 150 feet north to display Corallian at about 500 feet on the upthrown side and the top of the Upper Greensand above 650 feet on the downthrown side. The base of the Gault rests on the Oxford Clay at about 400 feet under Pen Forest. This figure is supported by the amount the Cornbrash is displaced between Sunny Hill and Charlton Musgrove, north of Bayford, over a mile farther west.

Sunny Hill, north of Bayford, is formed of Cornbrash dipping 11° SE., and Forest Marble limestones have been found, low down on a shelf, on its northern slope. However, augering and remarks from the farmer show that soapy khaki Oxford Clay suddenly comes in again south of the stream.

Again local relief shows that a fault downthrowing north about 150 feet is required to drop the Cornbrash below the present surface east of Charlton Musgrove from the crest of Sunny Hill.

In several places west of Penselwood Hill, it is not always possible to disentangle genuine tectonic movement from later superficial cambering and landslips. However, the presence of the tufa spring and the relatively sharp and straight northern edge of the small sandy Upper Cornbrash inlier, west of Leigh Common, all point to a westward continuation of the Mere Fault, across the intervening Oxford Clay outcrop, to link up with the undoubted displacement of the Cornbrash north of Bayford, including Sunny Hill.

Finally, there remains the position of the Coneygore Hill–Stoke Trister Corallian outlier and the approach of Corallian and Cornbrash to within thirty yards of each other south-east of Bayford. Both oolitic formations show dips of 9–12° SE. in quarries, both on Sunny Hill and around Stoke Trister. Such a dip, seen in both formations, cannot bring in more than 250 feet of Oxford Clay below the Corallian under Coneygore Hill and the ridge north-east of it. Elsewhere within the Blackmore Vale, south of Wincanton, the Oxford Clay outcrop is up to three miles wide, and in the Westbury boring, sixteen miles to the north-east, the Oxford Clay was found to be 600 feet thick (Arkell, 1933). It is suggested that there is a fault, trending south-west to north-east, under Leigh Common, cutting out an unknown thickness of Oxford Clay and which could have a downthrow south-east of up to 250 feet (see Fig. 4, Section 6). The Old Series one-inch map (Sheet No. 18, 1st edition, 1850) marked such a fault bounding the north-west side of Coneygore Hill, but as the Corallian sequence appears intact along the north-west side of the Corallian outlier it is inferred that the fault lies undetected within the Oxford Clay. If it continues north-eastwards it must scissor across the Mere Fault somewhere near the tufa spring.

West of Coneygore Hill there are puzzling dips within the Cornbrash even closer to the base of the Corallian. Mr. House in April 1954 recorded a higher dip south in Upper Cornbrash, than that of 10° at a point marked on the old series map. To the west of it there is an isolated exposure of Lower Cornbrash. These are shown on Fig. 2, but there has been no opportunity for either of us to unravel what is clearly a complicated piece of mapping in detail; and there are problems here requiring further study.

(b) Interpretations and Correlations

(*i*) *The Mere Fault compared with other Tertiary Faults in South Dorset and the Hog's Back Faults in Surrey.* The characteristic feature of all the Tertiary folds that have rippled the Upper Mesozoic rocks of southern England is that the northern limbs are usually steeper than the southern, pointing to movement from the south. Where folding of Tertiary date passes into actual fractures, as it does along the Poxwell–Ridgeway fault zone in south Dorset, the Mere Fault, thirty miles to the north, and the Hog's Back strike faults in Surrey, the three faults have these features in common:

1. All three Tertiary faults downthrow north.
2. The three faults have been found to be steeply haded reversed faults with southward hade.
3. The Poxwell–Ridgeway Faults certainly are, and the Mere Fault may be, associated with earlier Intra-Cretaceous disturbances. The Hog's Back fault is strongly suggested also to be associated with earlier disturbances but it cannot be proved (Arkell, 1939).
4. The three faults possess throws which vary but range up to 600 feet.

In three aspects the Mere Fault differs from its counterparts in south Dorset and Surrey.

1. The absence of drag along the Mere Fault on the downthrown side is noticeable. Even the eight-foot thick Warminster Greensand laps right up to the Mere Fault between Dead Maid Quarry and the Corallian Hill in Zeals Park.
2. The Mere Fault does not appear to be associated with monoclinal folding, west of West Knoyle, whereas the Dorset faults and the Hog's Back faults fracture beds already tilted in part, though drag dips take the place of monoclinal dips west of Bincombe Tunnel in Dorset (Arkell, 1947a).
3. The Hog's Back strike faults and to a lesser extent the Ridgeway fault zone in Dorset have dip faults associated with them. No such dip faults have been found, so far, along the Mere Fault. No surface features suggest any and the augering results point in the same direction. The possible oblique fault at West Knoyle is a special case and is referred to in more detail below.

The Poxwell–Ridgeway Fault displays reversed movement of the older Intra-Cretaceous Abbotsbury Fault (Arkell, 1947a, 274, and Mottram & House, 1956), but no such reversed movement has been detected along the Mere Fault. The disposition of the Cornbrash between Bayford and north of Wincanton does not suggest any, but it remains to be worked out whether the Mere Fault, in its continuation westward into the Lower Oolite country west of Wincanton, is associated with an eastward prolongation of the faults at Sparkford, where a south throwing fault brings Lias against the Rhaetic inlier of Sparkford and Camel Hill (Kellaway & Wilson, 1941). Indeed such an association between south throwing faults and north throwing faults, the latter of presumed Tertiary Age, has been amply proved by M. R. House in the Weymouth peninsula.[1] This writer is now of the opinion that the Mere Fault, downthrowing north, and the Sparkford Fault, downthrowing south, are closely associated and that mapping along an undoubted fault zone between Wincanton and Sparkford Hill would settle the question.

(*ii*) *The Eastward Continuation of the Mere Fault*. It would be expected that the Mere Fault, with its maximum throw of at least 600 feet at West Knoyle, should not suddenly terminate on reaching the Chalk east of the church. Yet the inferred presence of the oblique tear fault suggests that the throw is very much reduced and that east of the Mere Fault's disappearance the relief from the earth movement is taken up by monoclinal folding. The high dip of 55° N. seen in the Upper Greensand, east of the farm buildings, may be a drag dip in part, but farther south-eastwards, towards Upton and Milton, the strike trend is away from the line of the Mere Fault produced eastwards, and the dips seen are probably monoclinal dips. Fig. 5 shows the outcrops at West Knoyle.

As F. H. Edmunds has described the Hog's Back faults (Dines & Edmunds, 1927) and has pointed the way concerning the Mere Fault (Edmunds, 1938), the writer referred the question of the fault's disappearance at West Knoyle to him. He regards the structure as follows (*In litt*. 25 May 1951): 'Two asymmetric folds to commence rising *en echelon*, really one fold with a twist. As pressure increases, the tension in the zone A–B (Fig. 3), which arises is released either by twisting of the beds as sometimes occurs in Surrey or by a fault. Assuming the latter, the fold A and the fold B continue to grow: and also the tear fault A–B. But fold A and B may, due to pressure, also go with thrust faults or rather a combined thrust fault, and one may be thrust forward to a greater degree than the other in which case contemporaneous thrust and tear faulting arises.' That is what Dines & Edmunds (1927) envisage for the Hog's Back, and from this author's description, F. H. Edmunds considers the faulting at West Knoyle is essentially similar. The Mere Fault is taken up by monoclinal folding east of West Knoyle, the two linked by the inferred oblique fault and which can be demonstrated on the ground.

The Albian strike ridge is set about 100 yards north of the Mere Fault, and appears to be part of the easterly block that has been driven farther north in yielding by folding than by the fracturing along the Mere Fault. So on present available evidence at West Knoyle, it appears that the Mere Fault is rapidly replaced by steep monoclinal folding, the latter probably accompanied by undetected slide planes. F. H. Edmunds concluded by remarking that 'it is interesting to find the fault transferred, as far as surface indications are concerned, to folding the Chalk (and Albian) which is not unexpected'.

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[1] Personal communications.

An alternative explanation of the West Knoyle structure and the continuation eastward of the Mere Fault has been given by M. R. House when he read the typescript, in 1957, based on his work on the Weymouth peninsula. He suggests that east of West Knoyle the steep reversed fault plane (dipping south at a high angle) is bent flatter by the undoubted monoclinal folding farther east. Thus with the regional easterly tilt this would bring a flatter lying fault plane below the present ground surface east of West Knoyle. Thus the fault trace could run into the Chalk down north of the Upper Greensand Cleeve Hill as shown on the sketch-map (Fig. 2). This suggestion by Dr. House is based on the bending of the fault plane suggested by evidence available along the Ridgeway–Poxwell fault zone in south Dorset (Arkell, 1947a, and House, *in litt.*). As part of the Mere Fault trace lies beneath the fan gravels south of Manor Farm at West Knoyle, the interpretation can only remain a matter of opinion, given the available evidence. In the down, north-east of the farm buildings, small exposures in chalk can support either hypothesis. Further information is needed before a definite conclusion can be reached as to how the Mere Fault and the folding are linked and whether the fault does continue as one fault or as many slide planes within the Chalk towards the east (see also Fig. 4, Section 3).

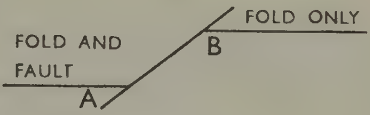

Fig. 6

Nevertheless, other authors (Jukes-Browne, 1900, map, p. 248, and Wooldridge & Linton, 1939) have drawn the Mere Fault as continuing within the long stretch of Chalk country east of West Knoyle for twelve and a half miles to link up with another fault downthrowing north which appears on the one-inch sheet map 298 as starting about a mile north-east of Dinton Church (41/020328).

This fault is marked as trending across the Nadder Valley at Barford St. Martin to disappear again between Hoop Side at the east end of the Chalk escarpment and south-east of Burcombe. The presence of this fault along the northern flank of the Wardour fold at its eastern extremity is not conclusive on present available evidence north of the River Nadder. It is perhaps significant that the higher dips seen in the chalk pits by Jukes-Browne occur near this fault. In particular, the pit three-quarters of a mile north-east of Dinton Church (41/016336) displayed an increase of dip from 20–45° between its two ends, a distance of 100 yards, and these may be drag dips (Jukes-Browne, *in* Reid, 1903, 49). In another pit in Middle Chalk, a half mile north-east of Baverstock Church, a dip of 33° (41/036323) ecorded by Jukes-Browne *in* Reid, 1903, is on the north side of the Chalk Marl vale which is narrower than the same strike vale is farther west. The vriter found that the dip slope of Albian chert beds steepens markedly, ust south of the Chalk Marl outcrop, near the northern edge of Crowdell's Copse, one-third of a mile north-east of Baverstock Church (41/033321), et he found an ammonite, *Schloenbachia varians* (J. Sow.), within a few ards of the chert debris, suggesting that the Cenomanian base is not cut ut by faulting. The Greensand dip, 15° in the neighbouring railway cutting, loes not appear to show a steepening towards the fault line. East of Hoop Side, south-east of Barford St. Martin and south of Burcombe, this roblem needs further investigation.

It appears doubtful therefore if the Mere Fault can be traced east of West Knoyle, towards Baverstock and Barford St. Martin, without areful zonal mapping of the chalk in the long stretch of intervening lownland. Judging from Dorset evidence (Arkell, 1947a, and Mottram & House, 1956) there is no theoretical reason why fractures should not ccur, expressed at the surface entirely within chalk, but it does seem likely hat the throw of these fractures cannot amount to anything like the 600 eet at West Knoyle, where the Mere Fault runs into the Chalk.

3. CONCLUSIONS

(a) The Wardour Fold

The Wardour Fold is Tertiary in age, involving Upper Jurassic and Cretaceous rocks, and the anticlinal axis is at least ten miles long, extendng from Barford St. Martin to East Knoyle.

The fold is asymmetrical, dips on the southern limbs range up to 5°, on he northern limb dips usually range up to 10–20° but locally exceed 40°, eaching 55° at West Knoyle, and has an amplitude a little over 1200 feet. A ubsidiary periclinal roll, riding on the northern limb of the fold, is liscernible north of East Knoyle.

Local rolls within the Jurassic rocks, in the Vale of Wardour, are ransgressed by local Aptian and Albian, thereby indicating that these mall folds are Intra-Cretaceous in age.

(b) The Mere Fault

The Mere Fault is a reversed fault, downthrowing north, bringing Jurassic against Cretaceous rocks, the downthrow decreasing from east to west over a distance of nine miles from above 600 feet at West Knoyle and Charnage, to 400 feet at Mere and to over 150 feet at Penselwood and Bayford. No reversed movement, along older fracture lines, has been detected along the Mere Fault.

The fault is associated with local disturbances which affect the Jurassic rocks on the upthrown side. These could either be local ruckles on the overdriven upthrown side or older? Intra-Cretaceous periclinal domes that the later Mere Fault has followed.

As Cretaceous beds are involved, the Mere Fault is held to be of Tertiary Age. The Mere Fault is similar to the Hog's Back strike faults and also to parts of the Dorset Poxwell–Ridgeway faults.

The Leigh Common fault only affects Jurassic rocks and may be of Intra-Cretaceous age.

The Mere Fault can be traced into the Jurassic country near Wincanton (and may extend farther west), only affecting Jurassic rocks.

East of West Knoyle the Mere Fault appears to be replaced by monoclinal folding.

ACKNOWLEDGMENTS

This writer wishes to thank a number of people who have helped him at various times since work was begun in 1951. Mr. F. H. Edmunds encouraged the author initially and subsequently gave information that could only have come from a native of Mere. Both the late Dr. W. J. Arkell, F.R.S., who was taken over some of the ground in August 1951, and Dr. P. E. Kent, Senior Geologist of the D'Arcy Exploration Company (as it then was) who likewise was taken over the area in May 1954, have given advice during discussion. Further help and information have come from Messrs. Colin Forbes, James Perry, Michael House, Michael Hancock and S. W. Hester. These last two helped in arranging for typescripts to be prepared.

It should be said that most of the work on which this paper is based was done in 1951–2, while resident at East Knoyle. Further time was spent in 1954 before the writer went to Uganda. A few extra details on the Lady Down outcrops were added during home leave in May 1957. The palaeontological records, up to 1954, have been included in the 1954 Whitsun Field Meeting report (Mottram, 1956). The bulk of the material was written and revised during 1958 and 1959. Finally it must be said that there is a great deal of research still to be done in this area and this paper is no more than a contribution to our knowledge on the Geology of the Mere Fault and the Vale of Wardour Anticline.

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The Highest Lower Old Red Sandstone of Brown Clee Hill, Shropshire

by J. R. L. ALLEN

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BSTRACT: The unfossiliferous Lower Old Red Sandstone beds above the Ditton eries at Brown Clee Hill, Shropshire, are chiefly sandstones with subordinate silt-tones, conglomerates, limestones and dolomitic rocks. They are divided into the bdon Series (Hillside Dolomitic Formation, 12 feet; Nordybank Sandstone Forma-ion, 200 feet; Abdon Limestone Formation, 17 feet) overlain by the Woodbank eries (Clee Sandstone Formation, 550 feet; Monkeys Fold Sandstone Formation, 20 feet). The beds are folded into a shallow asymmetrical syncline of caledonoid trend nd are overlain unconformably by eastward-dipping Coal Measures. The Abdon eries is here correlated with the upper part (Upper Dittonian) of the Red Marls of outh Wales and the Woodbank Series with the Senni Beds (Breconian). The Abdon eries and Woodbank Series were formerly known as the Clee Series (here discon-nued), which was correlated with the Senni Beds alone.

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1. INTRODUCTION

OCKS OF THE Lower and Upper divisions of the Old Red Sandstone crop ut widely in South Wales and the Welsh Borderland. The Upper Old ed Sandstone, passing into the Lower Carboniferous, is less than 1200 eet thick and lies with a depositional break on beds of the Lower division, nd at some places on still older rocks. This unconformity can now be stablished (Thomas, 1960) between the Plateau Beds, of proved Upper ld Red Sandstone age, and the Brownstones, of inferred Lower Old Red andstone age, at several localities in the Carmarthenshire Vans. However, he break occurs only as a disconformity in the Brecon Beacons. These reas, intermediate between Pembrokeshire and the Welsh Borderland, are he only ones where the existence of the break between Upper and Lower ivisions of the Old Red Sandstone had hitherto seemed uncertain or nprobable (Robertson, 1932; Pringle & George, 1948, 49).

The Lower Old Red Sandstone reaches about 3500 feet in the Welsh Borderland and follows with but a small break from the Silurian (Stamp, 1923). South-westwards to South Wales and north-westwards to Anglesey an unconformity is developed and the Lower Old Red Sandstone, 7000 feet thick in Carmarthenshire, steps across the Lower Palaeozoic to reach Pre-Cambrian beds (Greenly, 1919; Straw, 1930).

The Clee Hills of Shropshire rise from Corvedale on the west and the Wyre Forest on the east. They are dominated by the prominences of Brown Clee Hill in the north and Titterstone Clee Hill in the south. Each hill is carved mainly from Lower Old Red Sandstone rocks but is capped by Carboniferous beds. At Titterstone Clee Hill the Upper Old Red Sandstone crops out, resting unconformably on the gently folded Lower division. Brown Clee Hill alone provides good sections in the highest of the Lower Old Red Sandstone beds.

Conybeare & Phillips (1822) and J. Wright (1832) long ago recorded from Shropshire red sandstones, mudstones and limestones of the Old Red Sandstone. The Old Red Sandstone of the Clee Hills was early described by R. Wright (1833) and Murchison (1839) both of whom observed that on Brown Clee Hill two prominent limestones occurred high up in the succession. Between 1850 and 1868 Aveline, Hull and Ramsay mapped the hills for the Geological Survey, and Pocock (*in* Whitehead & Pocock, 1947) later described the north-eastern sector. Recently ground in the west and south has been remapped (Wilson, 1957).

Modern knowledge of the Old Red Sandstone of the Clee Hills stems from the work of Wickham King (1925, 1934). He established a detailed lithological succession in the area and attempted to use the vertebrate faunas, known for more than a century, in the development of a scheme of 'divisions' and 'stages'. White & Toombs (1948) and White (1950, 1956) revised this scheme, developing a more comprehensive zonal classification based on careful palaeontological collecting in the Welsh Borderland and South Wales. Dineley (1951) recently described the lower part of the Lower Old Red Sandstone succession of the Clee Hills and gave an important place to the rock types and depositional structures found. A general account of the area has been given by Ball & Dineley (1952) who recognised the following Old Red Sandstone succession.

UPPER OLD RED SANDSTONE	
Farlow Sandstones Series (Farlovian)	about 400 ft.
Unconformity	
LOWER OLD RED SANDSTONE	
Clee Series (post-Dittonian)	about 900 ft.
Ditton Series (Dittonian)	1000–1200 ft.
Downton Series (Downtonian), with Ludlow Bone Bed at base.	about 1000 ft.
SILURIAN	

The Clee Series of these authors passes up from the Ditton Series and occurs chiefly at Brown Clee Hill. At Titterstone Clee Hill a small and ill-exposed outcrop of the lowest beds is known. D. L. Dineley and D. W. Gossage find evidence for the limited occurrence of the group in the Cleobury Mortimer area (Dineley, 1959, *in litt.*).

Perhaps because of its lack of fossils the Clee Series is poorly known. Yet it includes the youngest Lower Old Red Sandstone beds in the Welsh Borderland, lying east of the Church Stretton axis and more than thirty miles away from any other strata of the same age. Recent work by the writer at Brown Clee Hill has shown that the Clee Series bears a wide variety of grains and pebbles, many of which are fossiliferous. Their study, combined with research on the rock types, depositional structures and rock associations of the Clee Series, sharply illuminates the deposition of the Lower Old Red Sandstone and the evolution of Wales and the Welsh Borderland during the earliest phases of the Middle Old Red Sandstone (Caledonian) orogeny. This work will shortly be reported. The present paper describes the detailed succession of the Clee Series, filling an important gap in knowledge and providing the essential framework to the sedimentological studies.

2. NOMENCLATURE OF THE HIGHEST LOWER OLD RED SANDSTONE

Mapping has shown that the Clee Series of Ball & Dineley (1952, 211) is sufficiently diverse to warrant a finer stratigraphical division than had previously been attempted. New lithological divisions are introduced and named in this paper and the term ‘Clee Series’ is abandoned. A revised correlation of the beds with the South Wales succession is suggested, and this is embodied with the new stratigraphical terms in Table I.

Two prominent limestone bands, the ‘Abdon Limestones’, were recognised by Ball & Dineley within their ‘Clee Series’. The base of the lower limestone was taken as the base of the ‘Clee Series’ and the top of the Ditton Series. Examination shows that the 200 feet of beds between the ‘Abdon Limestones’ differ markedly from the rocks above and resemble more nearly the Ditton Series beneath. Above the higher limestone green and red fine-grained to coarse-grained sandstones crop out. The chief rocks between the two limestones are red very fine-grained sandstones and siltstones. Thus, together with the limestones, the beds between the ‘Abdon Limestones’ can be separately distinguished from the red and green sandstones above, here termed the Woodbank Series. Within this series two formations can be recognised. They are the Clee Sandstone Formation (green sandstones and red siltstones) and the Monkeys Fold Sandstone Formation (green and red sandstones).

TABLE I. **Lower Old Red Sandstone stratigraphy at Brown Clee Hill and suggested correlation with South Wales**

		SOUTH WALES	BROWN CLEE HILL (this paper)
LOWER OLD RED SANDSTONE	BRECONIAN STAGE	BROWNSTONES	
		SENNI BEDS	WOODBANK SERIES Monkeys Fold Sandstone Formation, 0–220 feet Clee Sandstone Formation, 500–550 feet
	DITTONIAN STAGE	RED MARLS (higher part)	ABDON SERIES Abdon Limestone Formation, 15–17 feet Nordybank Sandstone Formation, 180–200 feet Hillside Dolomitic Formation, 8–12 feet
			DITTON SERIES (sense of Ball & Dineley, 1952)

The affinities of the beds between and including the 'Abdon Limestones' may now be examined. The Ditton Series presents a monotonous succession in which the same rock types are often regularly repeated. With the exception of one of the 'Abdon Limestones' there is no band sufficiently widespread or distinctive to form a practicable upper boundary to the group. Furthermore, the Ditton Series and the succession between the limestones are similar in that both consist chiefly of red siltstones and very fine-grained sandstones. Nevertheless, the two formations can be separated. Sandstones total 50–90% of the rocks between the 'Abdon Limestones' whereas in the Ditton Series they rarely aggregate more than about 40%. The strata between and including the limestones are therefore recognised as a separate lithological group, here named the Abdon Series. The upper lithological boundary of the Ditton Series remains unchanged at the base of the lower limestone band.

The 'Abdon Limestones' can be distinguished from each other more clearly. The lower contains abundant dolomite (10–60%) whereas dolomite is more rare (3–16%) in the upper bed. The two bands likewise differ in their field appearence. The lower band is accordingly designated as the Hillside Dolomitic Formation and the upper band the Abdon Limestone Formation. The red strata between are named the Nordybank Sandstone Formation.

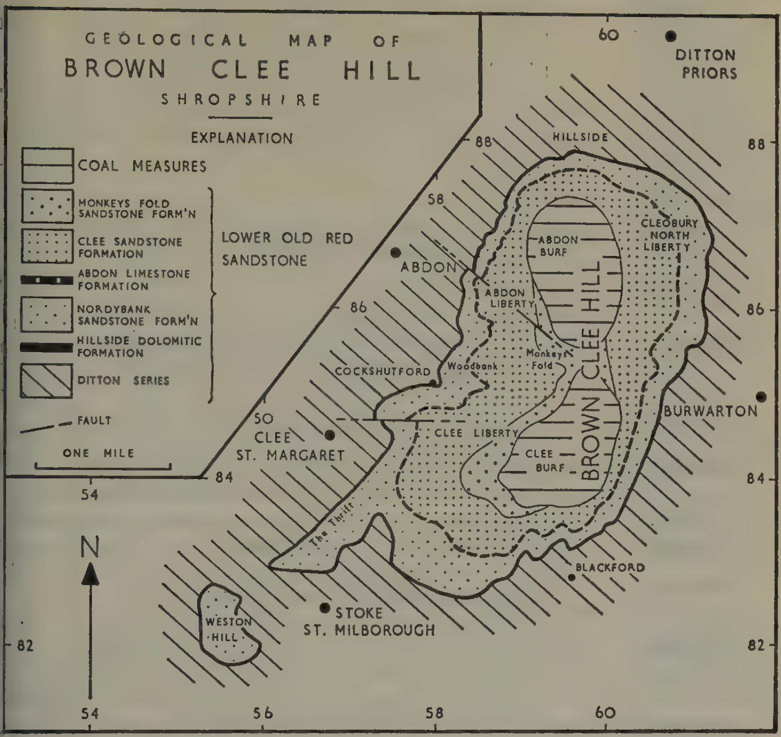

Fig. 1. A simplified geological map of Brown Clee Hill, Shropshire

Fig. 1 shows the outcrops of these formations on Brown Clee Hill, Fig. 2 the generalised succession of beds.

3. THE ABDON SERIES

(a) Hillside Dolomitic Formation

This formation, between eight and twelve feet thick, is dominated by calcite-dolomite rocks of varied grade accompanied by subordinate sandstones and siltstones. The beds are chiefly exposed in quarries scattered westwards and southwards along the outcrop from near Hillside (604876)[1] to Abdon Liberty (583857). The outcrop, rarely exposed, extends thence to Nordybank and The Thrift. The formation is seen neither in the outlier of Weston Hill nor between Stoke St. Milborough and Blackford. On the east of Brown Clee Hill the group is unexposed but its position was marked at the foot of steep slopes locally underlain by red sandstone debris from

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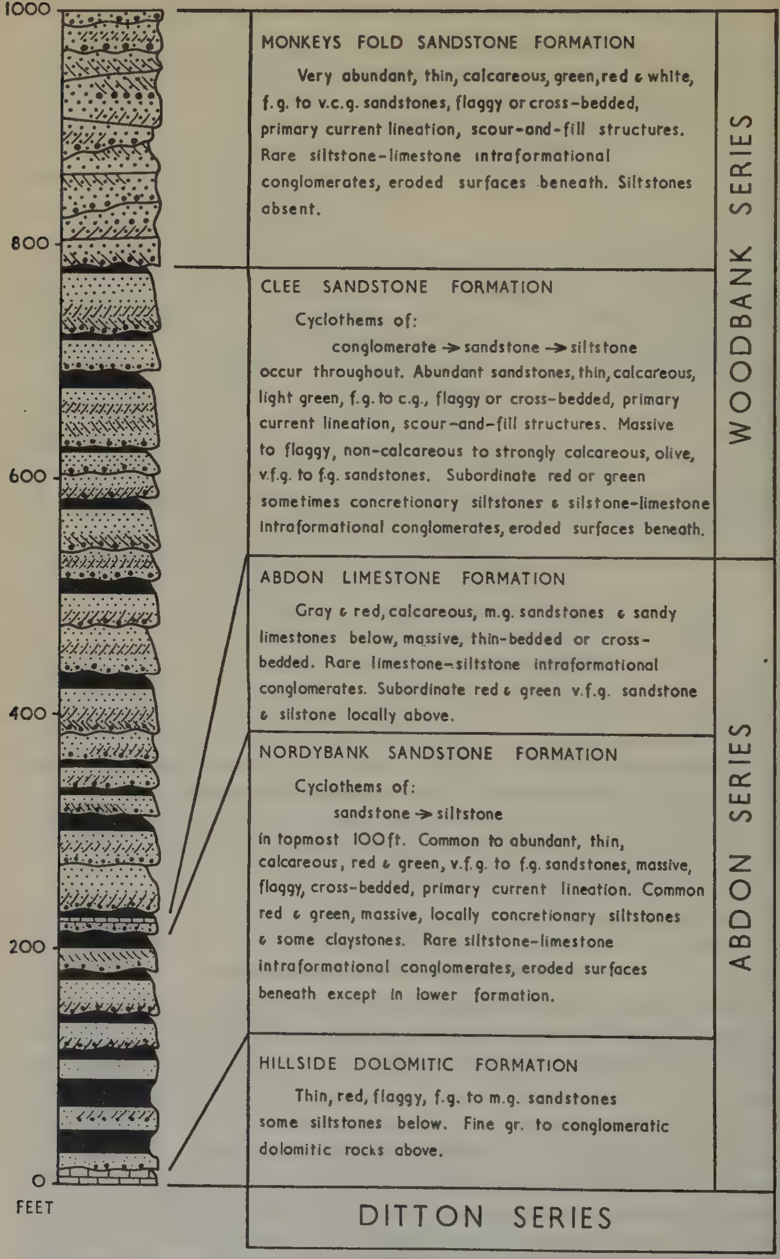
[1] The numbers enclosed in brackets are those of the National Grid Reference system, 100 km. square 32 (SO).



MONKEYS FOLD SANDSTONE FORMATION

Very abundant, thin, calcareous, green, red & white, f.g. to v.c.g. sandstones, flaggy or cross-bedded, primary current lineation, scour-and-fill structures. Rare siltstone-limestone intraformational conglomerates, eroded surfaces beneath. Siltstones absent.

CLEE SANDSTONE FORMATION

Cyclothems of:

conglomerate → sandstone → siltstone

occur throughout. Abundant sandstones, thin, calcareous, light green, f.g. to c.g., flaggy or cross-bedded, primary current lineation, scour-and-fill structures. Massive to flaggy, non-calcareous to strongly calcareous, olive, v.f.g. to f.g. sandstones. Subordinate red or green sometimes concretionary siltstones & silstone-limestone intraformational conglomerates, eroded surfaces beneath.

ABDON LIMESTONE FORMATION

Gray & red, calcareous, m.g. sandstones & sandy limestones below, massive, thin-bedded or cross-bedded. Rare limestone-siltstone intraformational conglomerates. Subordinate red & green v.f.g. sandstone & silstone locally above.

NORDYBANK SANDSTONE FORMATION

Cyclothems of:

sandstone → siltstone

in topmost 100 ft. Common to abundant, thin, calcareous, red & green, v.f.g. to f.g. sandstones, massive, flaggy, cross-bedded, primary current lineation. Common red & green, massive, locally concretionary siltstones & some claystones. Rare siltstone-limestone intraformational conglomerates, eroded surfaces beneath except in lower formation.

HILLSIDE DOLOMITIC FORMATION

Thin, red, flaggy, f.g. to m.g. sandstones some siltstones below. Fine gr. to conglomeratic dolomitic rocks above.

Fig. 2. The generalised succession of the Abdon Series and Woodbank Series. The grades and structures of the beds are indicated by conventional symbols

the Nordybank Sandstone Formation. Unfortunately, thick superficial deposits largely obscure the solid rocks in this area and the boundaries shown on the map are plotted with reservations.

The Hillside Dolomitic Formation is best exposed near Abdon and Blackford. In two quarries (583857, 583860) south-east of Abdon two feet of greyish-red, calcareous, fine-grained sandstones are overlain by hard, pinkish-grey to pinkish-white, thin to thick-bedded pebble and granule conglomerates totalling about eight feet. The well-rounded granules and pebbles are chiefly of argillaceous limestones and calcareous siltstones. Crystalline dolomite and calcite cement this debris and there is a little quartz sand. The top of the formation is not exposed here, but in a quarry (586869) east of Abdon red sandstones of the Nordybank Sandstone Formation rest unevenly on fine-grained dolomitic rocks of the group.

About ten feet of the Hillside Dolomitic Formation are exposed by the side of a track (596832) north of Blackford. About four feet of red sandstones and siltstones are here overlain by six feet of pale pink fine-grained dolomitic rocks. Sandstones of the Nordybank Sandstone Formation follow sharply, resting on an eroded surface.

(b) Nordybank Sandstone Formation

The Nordybank Sandstone Formation ranges between 180 and 200 feet thick and gives rise to the steep slopes at the foot of Brown Clee Hill. Red and green sandstones and siltstones are the predominant rocks whilst conglomerates are rare. They are well exposed in the stream gutters and crags extending from Hillside southwards to the brook at Cockshutford. Here, immediately beneath the Woodbank Series, red and green sandstones and siltstones crop out. A quarry at Nordybank shows similar rocks together with thin conglomerates. Features and occasional crags mark the outcrop from Cockshutford to The Thrift. Thin-bedded brown sandstones are well displayed in the lane (568836) joining Clee St. Margaret with Stoke St. Milborough. On Weston Hill sandstones and conglomerates are exposed. The beds are next seen in a gutter (591832) north of Blackford where thin-bedded red and green sandstones and siltstones outcrop intermittently. East of Brown Clee Hill the formation underlies steeper ground than the Ditton Series. Exposures are few but at Cleobury North Liberty (608871) and west of Burwarton (605855) red sandstones and siltstones are seen. The best exposures are at Hillside and Nordybank.

About forty-five feet of beds are exposed in a quarry (599878) at Hillside and begin within five feet of the formation base. Greyish-red and light greenish-grey sandstones in units up to five feet thick compose 57% of the length of the section. The sandstones are calcareous, massive to thin-bedded, and of very fine to fine grain. The coarsest are often cross-bedded with foresets dipping chiefly south-eastwards. Sometimes they show large scour-and-fill structures. Well-rounded pebbles of limestones and siltstones locally form impersistent conglomerates which aggregate 6% of the succession. They occur immediately below or interbedded with the sandstones. Units of massive or feebly laminated siltstones with some claystones aggregate 37%. They are red or green and commonly bear irregular calcareous concretions reaching six inches in diameter.

The quarry (577848) north of the earthworks on Nordybank exposes thirty feet of beds commencing about 100 feet above the formation base. Units of greyish-red and light green sandstones up to six feet thick total 90% of the succession. The sandstones are calcareous, laminated, thin-bedded or massive, sometimes cross-bedded and of very fine to fine grain. Many carry at or near their bases well-rounded pebbles of calcareous siltstones and argillaceous limestones. Similar pebbles form thin conglomerates totalling 3% of the section; 7% are massive, often concretionary, red and green siltstones in beds up to two feet thick. Normally they are overlain by sandstones, often separated from them by irregular eroded surfaces.

The beds at Cockshutford, Blackford, Nordybank and Burwarton are grouped into repeated cyclothems each up to twelve feet thick, consisting of sandstones graduating upwards into siltstones. Similar groupings were not identified in the lower part of the Nordybank Sandstone Formation exposed at Hillside and elsewhere.

(c) Abdon Limestone Formation

Red to grey calcareous sandstones grading to sandy limestones are dominant and locally pass up into red and green very fine-grained sandstones and siltstones. The group, between fifteen and seventeen feet thick, gives rise to a prominent bench on the western and southern slopes of Brown Clee Hill. The beds are exposed in quarries south of Clee Burf and in others extending from Cleobury North Liberty westwards and southwards to Abdon Liberty. Exposures are few on the east of the hill and there the formation rarely forms a strong feature.

The Abdon Limestone Formation is well exposed in crags (587866–588869) to the east of Abdon. The lowest beds are ten feet of greyish-red, thin-bedded or massive, sometimes cross-bedded, strongly calcareous sandstones and sandy limestones. Thin and impersistent conglomerates of limestone and siltstone pebbles in sandy matrices are also present. The sandy limestones pass up into five feet of mottled, brownish-grey, massive, splintery, fine-grained argillaceous limestones bearing a little quartz sand. Here and at many other localities the upper limestones are marked by structures possibly of algal origin. These structures consist of irregular, broken, sometimes crenulated parallel bands of very finely crystalline light and dark calcite. At several of the crags a red pebbly sandstone, taken as the lowest unit of the Woodbank Series, rests sharply on an irregular surface eroded into the topmost limestones. A conglomerate locally present at the base of this sandstone contains fragments of the Abdon Limestone Formation.

In the stream (586851) east of Cockshutford the group begins with twelve feet of ill-exposed, medium-grained, calcareous sandstones and sandy limestones. The highest beds, marked by (?) algal banding, rapidly pass up into three feet of green and red very fine-grained sandstones overlain by red and green siltstones aggregating two feet. A green sandstone of the Clee Sandstone Formation follows sharply on an irregular surface.

A nearby quarry (580848) exposes a comparable though shorter succession. Six feet of mottled red and grey, massive to rubbly, argillaceous to sandy limestones underlie three feet of dark red and green very fine-grained sandstones and siltstones. The topmost siltstone carries smooth nodules of splintery limestone. A thin conglomerate of siltstone and limestone pebbles and cobbles overlies the siltstone and rests on an uneven surface. Fifteen feet of green sandstone follow and are placed with the conglomerate in the Woodbank Series.

The spoil heaps of two quarries (596834, 592833) north of Blackford yield many fragments of red and grey sandy limestones, some marked by (?) algal banding. The beds themselves are very poorly exposed. Similar limestones from the Abdon Limestone Formation form a poor outcrop in the stream (604855) west of Burwarton.

4. THE WOODBANK SERIES

(a) Clee Sandstone Formation

The Clee Sandstone Formation, between 500 and 550 feet thick, consists of dominant green sandstones, subordinate red siltstones, and rare conglomerates and dolomitic rocks. Its beds form the main mass of Brown Clee Hill and about half of the succession above the Ditton Series. The formation is well exposed in the streams at Clee Liberty, north-east of Stoke St. Milborough, north of Blackford and west of Burwarton. Crags of sandstones, sometimes resting sharply on siltstones or passing up into them, mark the outcrop on the hillsides between. They are abundant between Abdon Liberty and Clee Liberty and south of Clee Burf but are rare on the east of the hill. The outcrop of the formation is normally broad but by the overstep of the Coal Measures south of Hillside and Clee Burf its width, together with the thickness of the beds, is much reduced. The relatively high dip at Clee Burf contributes also to the diminution in the width of outcrop.

Green, pebbly, medium-grained sandstones of the Clee Sandstone

Formation crop out in pits (590864) on Abdon Liberty. A nearby quarry (587861) exposes about thirty feet of beds commencing about 100 feet above the Abdon Limestone Formation. Very fine-grained to coarse-grained light green sandstones are the main rocks but conglomerates of limestone and siltstone detritus are common. Thick units of dark red siltstones occur near the floor and top of the quarry and are each overlain by locally conglomeratic sandstones.

About 450 feet of the group are excellently exposed in the stream (583848–587844) draining north-westwards across Clee Liberty. Interbedded sandstones, siltstones and conglomerates are first seen about seventy-five feet above the formation base. Light greenish-grey to greenish-grey, calcareous, medium-grained and rarely coarse-grained sandstones form units up to six feet thick and total 36% of the beds. Included with them are a few dusky yellow decemented sandstones of the same grades. The sandstones are chiefly thin-bedded and flaggy. Primary current lineation (Stokes, 1947) is typical, appearing widely over their bedding planes as faint parallel ridges and furrows. Less often the sandstones are cross-bedded, their foresets inclining between east and south. Many sandstone units carry scour-and-fill structures.

Thin and often lenticular conglomerates, grading upwards and laterally into the sandstones described, make up 5% of the succession. They consist of well-rounded pebbles, cobbles or boulders of red, green or yellow siltstones and calcareous siltstones. Pebbles of argillaceous limestones are often abundant. Fragments of other rock types are very sparing. Matrices of fine to coarse-grained sandstones bind these phenoclasts and are coarsest when cobbles or boulders are present. The conglomerates always rest on uneven eroded surfaces and normally overlie siltstones. The siltstone fragments of the conglomerates are usually indistinguishable from the interbedded argillaceous rocks. The conglomerates are of intraformational type.

Pale olive to light greenish-grey, non-calcareous to abundantly calcareous, very fine-grained to fine-grained sandstones aggregate 51% of the beds at Clee Liberty. These sandstones are massive to thin-bedded and flaggy, rarely cross-bedded or marked by primary current lineation. They carry abundant clay and grade from the coarser sandstones into massive, sometimes laminated, dusky red or pale olive siltstones composing 8% of the beds. The siltstones range to four feet in thickness and often bear calcareous concretions locally fusing to rubbly bands. Thin beds of dusky yellow, massive, argillaceous calcitic-dolomite and dolomite-rock crop out for five feet about 205 feet above the formation base.

The rocks described above are arranged in a cyclothem between two and thirty feet thick repeated continuously throughout the succession. In all thirty-two cyclothems were recognised at Clee Liberty. Each begins with an eroded surface on which rest conglomerates or pebbly sandstones grading through finer sandstones into siltstones.

The stream (585852–595853) between the Monkeys Fold and Cockshutford fully exposes the Clee Sandstone Formation, here measuring 500 feet in thickness. The predominant rocks are light green to olive, very fine-grained to coarse-grained, massive, thin-bedded or cross-bedded sandstones. Dark red and light green to olive siltstones are subordinate and conglomerates are rare. Many cyclothems were identified.

The same rock types in like proportions are exposed for 200 feet immediately beneath the Monkeys Fold Sandstone Formation in Batch Gutter (592834–592836) north of Blackford. The beds are again formed into cyclothems and thin yellow dolomite-rocks crop out about 515 feet above the base. The unexposed strata down to the base of the formation total about 330 feet.

Green sandstones, dark red siltstones and thin conglomerates are ill-exposed in the brook (580834–583857) north-east of Stoke St. Milborough and in the stream (600855–605855) draining towards Burwarton. Several cyclothems were identified at each locality.

(b) Monkeys Fold Sandstone Formation

This group reaches 220 feet in thickness at Clee Burf and consists of fine-grained to very coarse-grained sandstones with subordinate conglomerates. It is distinguished from the Clee Sandstone Formation from which it passes by the absence of siltstones and the generally coarser grain. The Monkeys Fold Sandstone Formation immediately underlies the unconformable Coal Measures and forms a narrow outcrop to the east of Abdon Liberty and to the north-west and south of Clee Burf. The outcrop is broadest at the Monkeys Fold but the beds are thickest south-west of Clee Burf. The group is absent on the east of Brown Clee Hill, the Coal Measures here resting on the Clee Sandstone Formation.

Crags (593858) on the hillside east of Abdon expose the lowest fifty feet of the formation. Pale red, greyish-red and light greenish-grey, medium-grained to very coarse-grained pebbly sandstones total 68% of the beds. The sandstones reach three feet in thickness and many are thin-bedded and flaggy, bearing primary current lineation. Others of similar thickness are cross-bedded, their foresets dipping south-eastwards. Scour-and-fill structures are common. Calcareous, thin-bedded, fine-grained sandstones aggregate 27% of the beds whilst thin lenticular conglomerates of limestone and siltstone pebbles or cobbles total 5%. Scattered in the conglomerates are pebbles of vein-quartz and various sedimentary and igneous rocks.

At the Monkeys Fold (594855), roughly seventy-five feet above the formation base, calcareous, very coarse-grained, pebbly cross-bedded sandstones are exposed. Red and green, medium-grained and coarse-grained sandstones also occur and are seen again in the brook (600855) west of Burwarton. Similar beds, reaching to about 100 feet from the base of the group, are exposed on the north-west shoulder of Clee Burf (586843–594850). Younger beds crop out to the south-west of Clee Burf (587840, 588839). They are chiefly pale red to white, pebbly, medium, coarse and very coarse-grained sandstones interbedded with conglomerates of vein-quartz and limestone pebbles.

5. STRUCTURE OF THE LOWER OLD RED SANDSTONE

A shallow asymmetrical syncline is formed by the Old Red Sandstone rocks at Brown Clee Hill. This downfold is flanked to the south-east by a broad anticline descending to a further syncline at Titterstone Clee Hill. Beds on the north-western limb of the Brown Clee Hill syncline dip at between 3° and 10° to the south-east and south-south-east. The opposite limb is more steeply inclined and at Blackford the Ditton Series, Abdon Series and Woodbank Series dip between 15° and 25° to the north-west. The axis of the syncline pitches gently north-eastwards and crosses the hill from Stoke St. Milborough to Cleobury North Liberty.

A fault trending south-eastwards breaks the north-western limb at Abdon Liberty and downthrows the Abdon Series about fifty feet to the south. The throw is twenty-five feet at the Monkeys Fold and farther south-eastwards the fault appears to die out. A second fault, downthrowing about thirty feet to the north, cuts the same limb at Nordybank but was not established far to the east or west. Minor faults occur in many of the stream sections.

The Coal Measures of Brown Clee Hill are pierced by a dolerite sill and dip at about 5° to the east or south-east. They form outliers at Abdon Burf and Clee Burf and lie across the bevelled edges of the folded Woodbank Series. South of Clee Burf the Measures step rapidly eastwards across the Monkeys Fold Sandstone Formation, finally resting on about 150 feet of the underlying division. The overstep northwards from Clee Burf to Abdon Burf is less rapid. Thus south of Clee Burf the Monkeys Fold Sandstone Formation is 220 feet thick whilst at the Monkeys Fold little more than seventy-five feet are preserved. South of Hillside the Coal Measures rest on the lowest 200 feet of the Clee Sandstone Formation.

6. AGE OF THE ABDON SERIES AND WOODBANK SERIES

An indigenous flora and fauna has not yet been reported from the Abdon Series or Woodbank Series despite their passage from the vertebrate-rich Ditton Series. However, their age can be suggested by comparing the succession at Brown Clee Hill with that of the Lower Old Red Sandstone of South Wales, where the vertebrate faunas range higher than in Shropshire.

The Senni Beds of South Wales yield *Rhinopteraspis dunensis* (*s.s.*) and overlie the Red Marls (sense of Strahan, 1899), bearing *Pteraspis* (*Protopteraspis*) *leathensis* and, at higher levels, *Pt.* (*Belgicaspis*) *crouchi* (White, 1950, 56–7). These species, together with the immediately younger *Pt.* (*Rhinopteraspis*) *leachi*, define the Dittonian stage (White, 1956, 5) to which the Ditton Series of the Clee Hills is assigned. Croft (1953, 431) defined the next highest stage, the Breconian (White, 1950, fig. 1, 53), on he occurrence of *Rh. dunensis* (*s.s.*) and also a distinctive flora. In the Breconian he placed (Croft, 1953, 429) the Senni Beds and Brownstones of South Wales and the 'Clee Series' of Shropshire, now divided and renamed he Abdon Series and the Woodbank Series.

Ball & Dineley (1952, 212) correlated the 'Clee Series' with the Senni Beds of South Wales. But the major lithological break at the top of the Abdon Series, formerly the lower part of the 'Clee Series', is indistinguishable from the important change in South Wales from the Red Marls to the enni Beds. Here red siltstones with very fine and fine-grained sandstones harply give way upwards to abundant green sandstones typically of medium grade. The correlation of the Woodbank Series with the Senni Beds (Breconian) and of the Abdon Series with the upper part (Upper Dittonian) of the Red Marls therefore seems more probable. Combining he floral and faunal evidence, Croft (1953, 431) equated the Breconian with the Siegenian and Emsian of the marine Devonian on the Continent. More recently White (1956, 5) and Schmidt (1959, table 2, 34) have limited s range from the Middle Siegenian to the Lower or Middle Emsian. The Upper Dittonian of Schmidt, correlated by him with the Lower Siegenian, onsists of the single zone of *Pt.* (*Rhinopteraspis*) *leachi*. This species occurs ot far below the base of the Abdon Series in the upper beds of the Ditton eries (Ball & Dineley, 1952, 213).

At Titterstone Clee Hill the Farlow Sandstones Series rests with a ronounced unconformity on the Downton Series and Ditton Series (King, 925, 386; Ball & Dineley, 1952, 211). King believed that Upper Old Red andstone beds also cropped out on Brown Clee Hill to the north. To this ivision he assigned 'calcareous concretionary green and green-purple andstones and fine yellow sandstones in red marls' that 'form the scarp on he south side of Clee Burf'. These must surely be the Woodbank Series for which strong evidence of a Lower Old Red Sandstone age has just been dduced. Moreover, the Upper Old Red Sandstone is now known to rest verywhere with either an unconformity or a disconformity on the Lower ivision. This relationship can be demonstrated at Titterstone Clee Hill ut a major break within the Old Red Sandstone cannot be found at rown Clee Hill. King's recognition of the Upper division at Brown Clee Hill is therefore difficult to uphold for only the Lower Old Red Sandstone an be shown to be present.

7. CONCLUSION

The mainly arenaceous rocks which conformably overlie the Ditton Series on Brown Clee Hill can be divided on the basis of their lithology into five formations. The Abdon Series includes the three lowest formations and is an upward extension of the facies of the Ditton Series but richer in sandstones and carbonate rocks. The two highest formations compose the Woodbank Series, much thicker than the Abdon Series and strikingly different from it in the colour and grade of the beds. Nevertheless, the Woodbank Series marks a further increase in the abundance and coarseness of the sandstones and, like the upper part of the Abdon Series, includes cyclothems.

The Abdon Series and Woodbank Series were folded along caledonoid lines before Coal Measures times and perhaps before the deposition of the Upper Old Red Sandstone. Their age cannot at the moment be decided by palaeontological means though their field relationships clearly place them within the Lower Old Red Sandstone. Correlation with the fossiliferous Lower Old Red Sandstone of South Wales, however, strongly suggests that the Abdon Series and the Woodbank Series are Upper Dittonian and Breconian respectively.

ACKNOWLEDGMENTS

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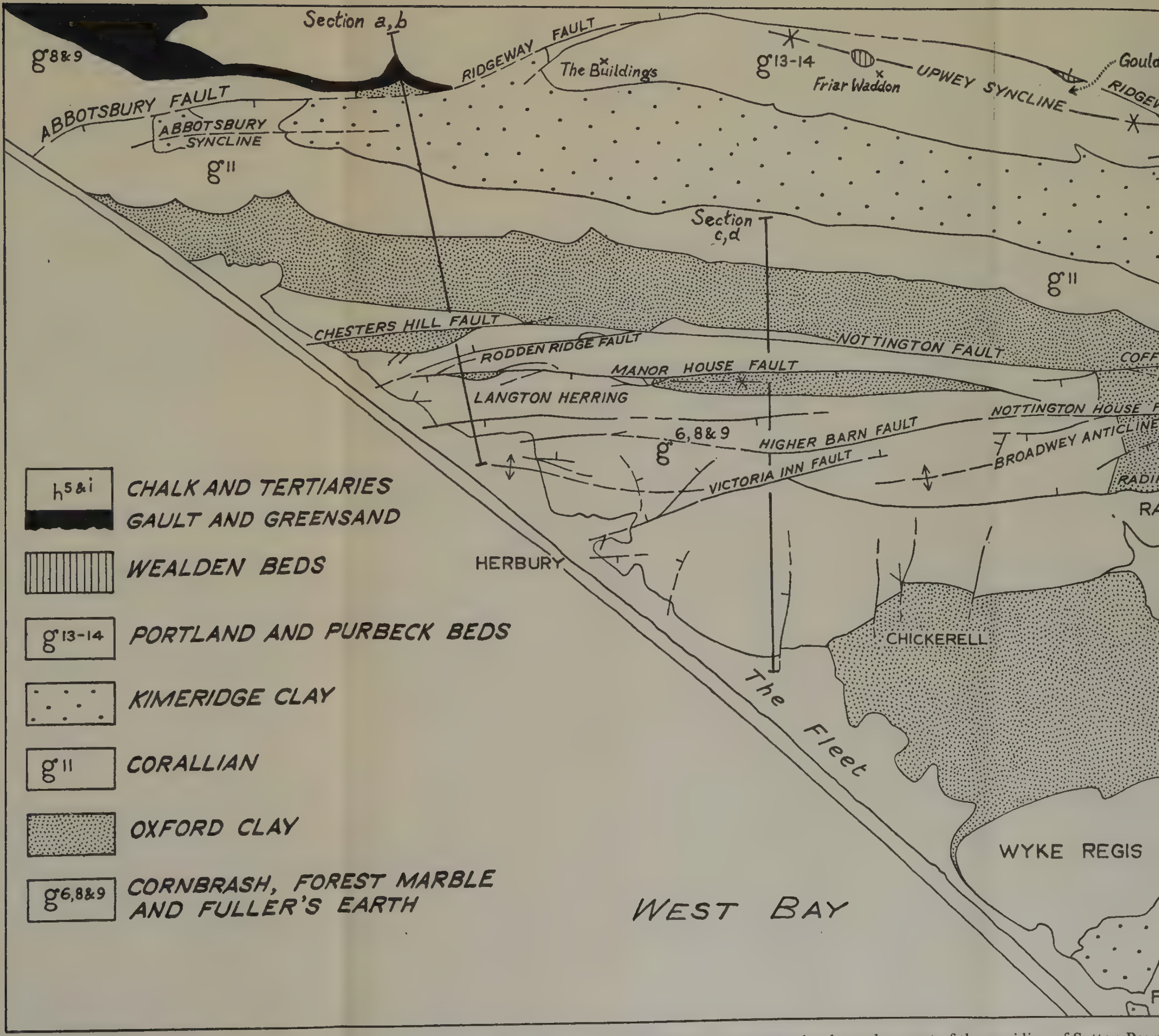

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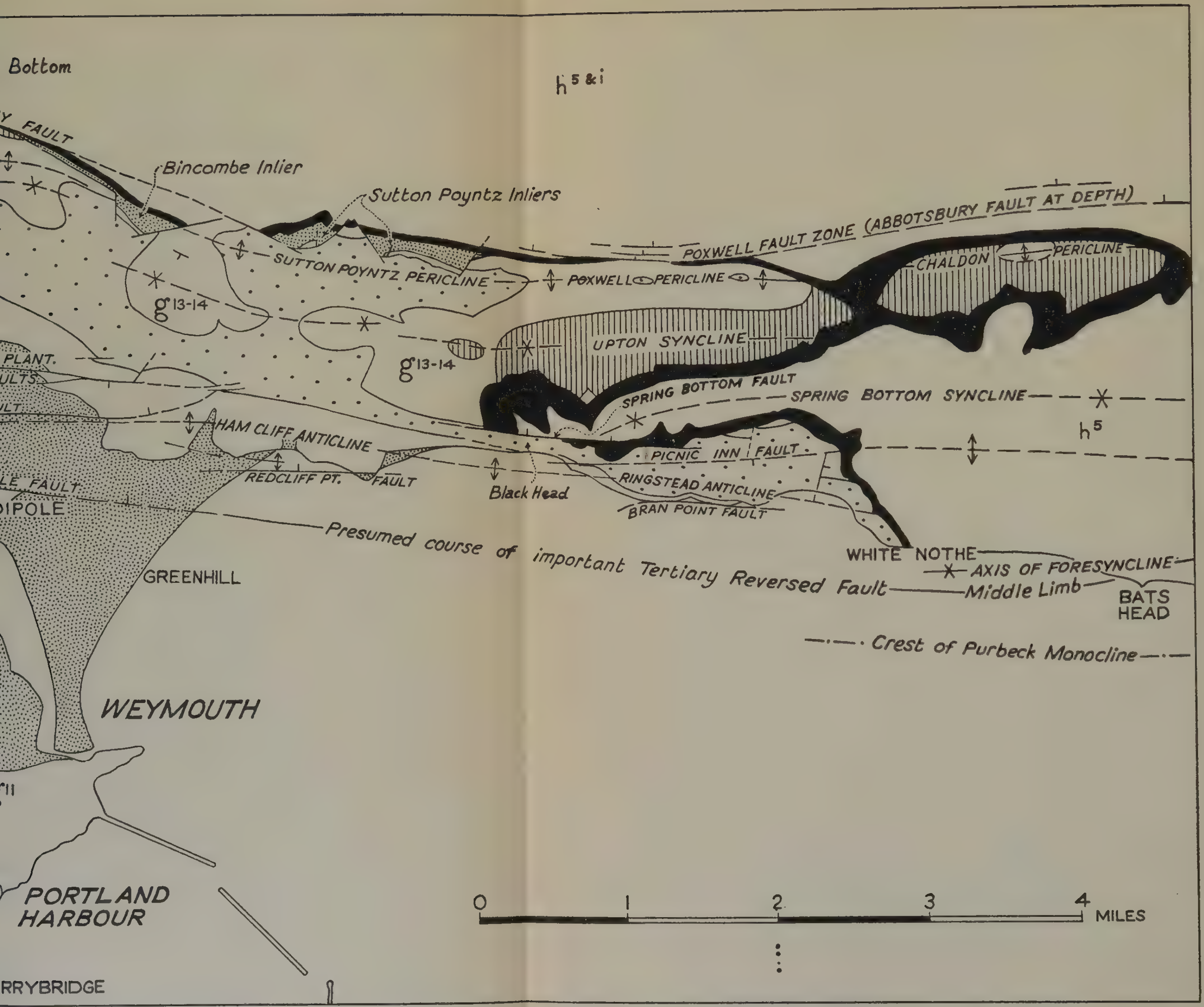
Fig. 1. A tectonic map of the Weymouth area. Based on mapping by the author west of the meridian of Sutton Poyr



and south of the Ridgeway Fault. Other lines from published maps by the Geological Survey, Arkell and Mottram & House

The Structure of the Weymouth Anticline

by M. R. HOUSE

Revised version received 21 December 1959

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ABSTRACT: An account of the structure of the Weymouth Anticline in south Dorset is based upon a new survey. Fuller's Earth crops out in the Anticline core from the Fleet shore to the banks of the Wey. The numerous strike faults which cut the core are described and some new structural details along the Ridgeway Fault belt are given. Borehole evidence obtained by the British Petroleum Company at Chaldon Herring is recorded and the evidence it provides on the Tertiary reactivation of an Intra-Cretaceous fault is discussed. An Intra-Cretaceous age for the Broadwey Anticline, as suggested by Strahan, is confirmed following the connection by mapping of the Broadwey with the Hamcliff Anticlines. The suggestion of Arkell that faults downthrowing north in the lowland are Tertiary in age and those downthrowing south are Intra-Cretaceous is supported. The westward disappearance of the Purbeck Monocline is discussed and it is suggested that the Radipole Fault may be the westward continuation of a reversed fault which replaces the middle limb of the Monocline near White Nothe.

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1. INTRODUCTION

THE RURAL LOWLAND between Weymouth, Abbotsbury and Upwey in south Dorset forms a natural geographic unit limited to the north by the chalk upland of Ridgeway, to the east by Weymouth Bay and to the south-west by the Chesil Beach and West Bay. The Isle of Portland, jutting out into the English Channel, forms a southern apex.

Through the centre of the Jurassic rocks of the lowland passes the east–west axis of the Weymouth Anticline (Fig. 1). The anticlinal structure was first noted by Conybeare and Phillips (1822, 182) and the earliest geological map was published in 1836 by Buckland & de la Beche. The faulted nature of the core was first well shown on the earliest one-inch Geological

Survey map of the area by Bristowe (1850, Old Series Sheets 16 and 17). Later Strahan surveyed the ground on the six-inch scale between 1888 and 1890 and the *Memoir* of the district by him (Strahan, 1898) and the accompanying one-inch maps (New Series Sheets 341 and 342) were for almost half a century the standard work. In 1932 the late Dr. W. J. Arkell began publishing a series of important papers on the area which culminated in an entirely revised *Survey Memoir* (Arkell *et al.*, 1947; second edition with addenda and corrigenda, 1952) and a new edition of the one-inch map (Sheet 341). The region around Abbotsbury is included on the Bridport Sheet (327) which was re-surveyed between 1931 and 1936 by officers of the Geological Survey and the map was reprinted in 1948. The *Memoir* for this region has recently been published (Wilson *et al.*, 1958). Most of the new work of the last thirty years has been concerned with stratigraphical problems or with the structures along the Ridgeway Fault Belt and the eastern region around Sutton Poyntz, Osmington, Poxwell and White Nothe. This paper deals mainly with the somewhat neglected area between Abbotsbury, Upwey and Weymouth.

The country between Ferrybridge, Wyke Regis, Abbotsbury and Sutton Poyntz, and north to the Ridgeway Fault, has been mapped on the six-inch scale and in parts on the twenty-five-inch scale. The bulk of the work was done between 1951 and 1956, but observations were made in preceding and subsequent years. By the time the mapping was completed it was obvious that the slight culminations along the crest of the anticline might be suitable places to test for oil in the Bridport Sands or Lias. The results of this work were therefore put at the disposal of the British Petroleum Company. The Chairman and Directors of the Company have allowed the publication here of the results of their investigations around Chaldon Herring, upon the structural interpretation of which the author was consulted. It should be made clear that the well records which are summarised later are completely the work of those acknowledged in the text. Although the borehole data were discussed with the B.P. geological staff, the structural interpretations which are given are entirely the responsibility of the writer. Mr. N. L. Falcon, Chief Geologist to the Company, kindly read the initial typescript.

2. STRATIGRAPHICAL NOTES ON THE MIDDLE JURASSIC ROCKS

A complete account of the stratigraphy of the Weymouth area was given by Arkell in the recent *Memoir* so that it is not necessary to give a summary here. The new mapping has provided some additional information on the Middle Jurassic rocks which may be summarised briefly.

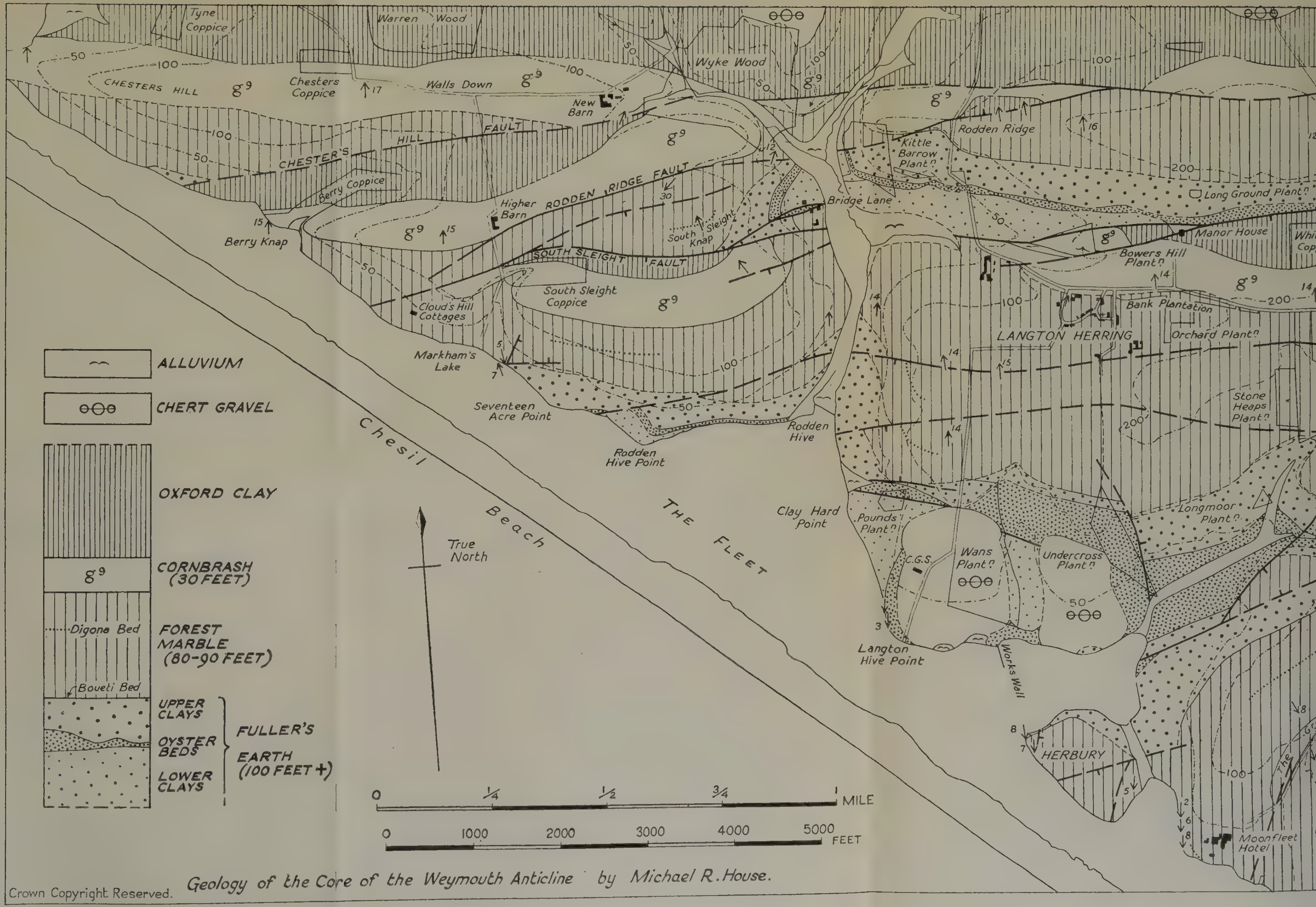

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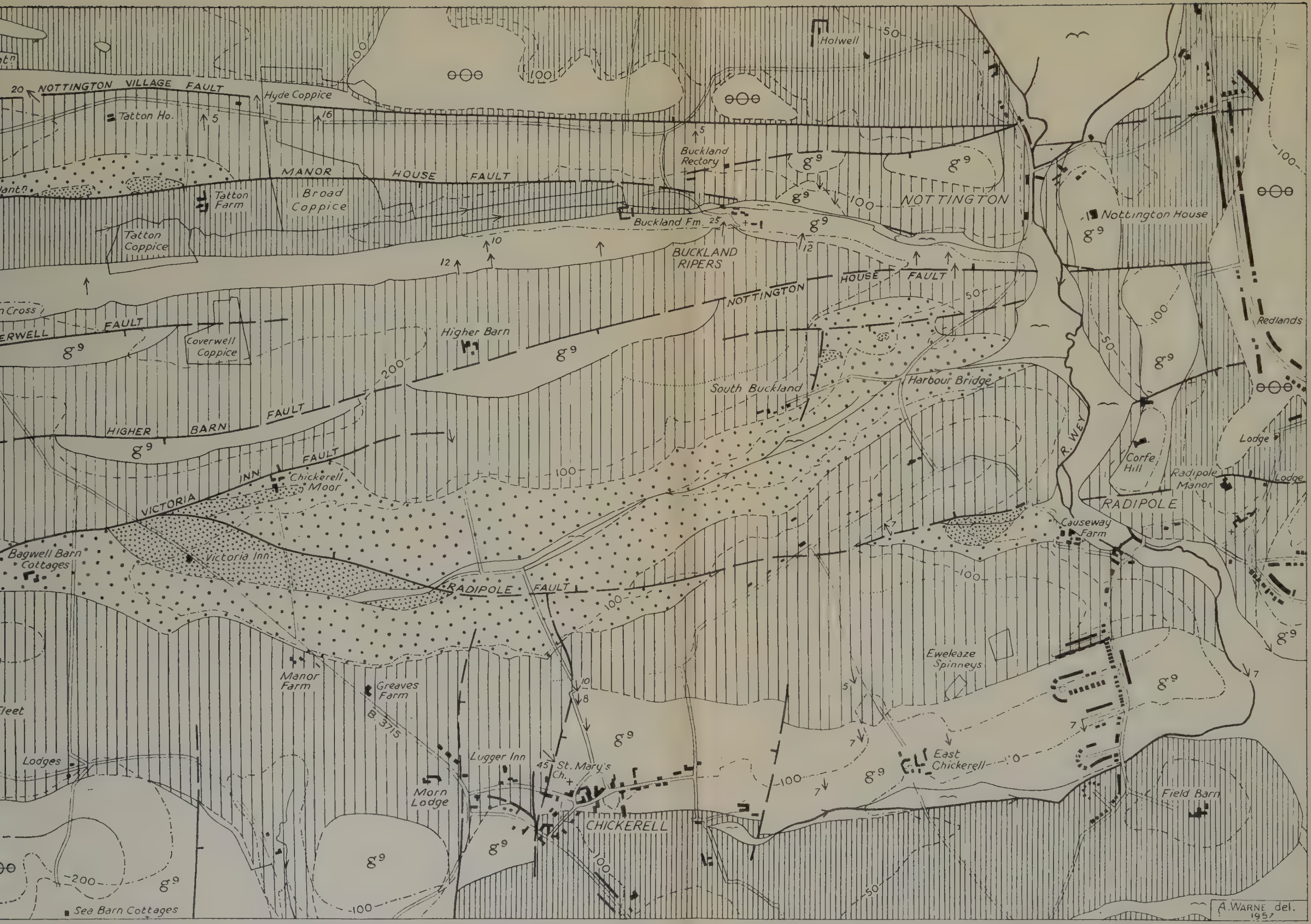
Fig. 2. The geology of the core of the Weymouth Anticline. Based on mapping by the author. Some dip readings made years ago in abandoned or overgrown quarries copie



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(a) Fuller's Earth

An account of the outcrop of these beds has already been published (House, 1957). The following succession of the Fuller's Earth has been determined for the Langton Herring area.

boueti Bed above (lowest Forest Marble)
Fuller's Earth (beds 5 and 6 seen at Herbury)

		ft.
6.	Fine white laminated limestone in three courses, the lowest thickening westward, separated by two beds of sandy shale	7–9
5.	Grey shale, seen to	15
4.	Clay, not exposed, about	15
3.	Oyster beds: lumachelle composed of *Ostrea hebridica* Forbes and var. *elongata* Dutertre, in places probably thinning out	0–*c*. 20
2.	Grey clay with well-preserved fauna at Rodden Hive Point (Arkell, 1940)	3
1.	Grey clay with white limestone nodules in the upper part, not well exposed; perhaps	50

(b) Forest Marble

The thickness of the Forest Marble estimated from the present survey is about ninety feet. Total measurements given in the past differ greatly. Damon gave an estimate of 500 feet (1884, 14) even after he had recognised, for the first time, the presence of Fuller's Earth along the Fleet shore. Woodward (1894, 338) gave a figure of eighty or ninety feet, and Strahan estimated the thickness as 130–140 feet near Moonfleet Hotel (Strahan, 1898, 7), but at that locality a fault is responsible for the high apparent thickness.

The *boueti* Bed, which is usually taken as the base of the Forest Marble (Woodward, 1894, 341; Welch, 1933, 40), forms a most valuable mapping horizon. In addition to an account of this bed given by Arkell (1947, 18) there are studies on the ostracods by Sylvester-Bradley (1948), and a long list of fossils found in the bed in north Dorset has been published by Fowler (1957).

Sylvester-Bradley (1957, 26) has recently given the name '*digona* Bed' to a fossiliferous horizon fifty-eight to sixty feet above the *boueti* Bed and near the top of the main series of flags which constitute the Middle Forest Marble. The horizon was first noted by Arkell (op. cit., 19) on Herbury peninsula. The bed has a very distinctive fauna and lithology and it has been mapped at several places inland which are marked on the accompanying map (Fig. 2). The fauna is characterised by *Ornithella digona* (J. Sowerby) and other brachiopods, mostly new, and also Bryozoa, especially *Berenicea* sp., sponges, *Apiocrinus* sp., *Nuculana* (*Dacromya*) *lacryma* (J. de C. Sowerby), *Chlamys* (*Radulopecten*) cf. *vagans* (J. de C. Sow.), *C.* (*R.*) *hemicostata* (Morris & Lycett), *Ostrea* cf. *hebridica* Forbes, *Nucula* (*Palaeonucula*) *waltoni* M. & L., *Lima* (*Plagiostoma*) *subcardiiformis*

Greppin, *Plicatula fistulosa* M. & L., *Oxytoma* cf. *costatum* (J. Sow.) and micromorphic gastropods.

The *digona* Bed seems distinguishable in several of the Forest Marble quarries described by Damon (1884, 14 *et seq.*). The Nottington Quarry described by him may be the one of which traces still remain 100 yards north of the Old Mill, Nottington, where the base of the Cornbrash lies twenty to thirty feet above the position where the *digona* Bed would be expected. This gives a total of about ninety feet for the Forest Marble as a whole, adding a few feet for the thickness of the *digona* Bed to the measured fifty-eight to sixty feet between the *boueti* Bed and the *digona* Bed.

3. THE GENERAL STRUCTURE

The structural crest of the Weymouth Anticline (Figs. 1 and 2) passes approximately due east from Clay Hard Point on the Fleet shore. Fuller's Earth is exposed in the core and younger rocks up to the Purbeck Beds are successively displayed to the north and south. The dips in the northern limb vary between 10° and 20° N., except where the Abbotsbury and Upwey synclines give southerly dips near the Ridgeway Fault where the Chalk is thrown down to the north against the Upper Jurassic rocks. The dips in the southern limb are more gentle, reaching 10° S. near the axis but gradually decreasing to 5° S. in the Wyke Regis Corallian Ridge and to 1½° SE. at the southern end of the Isle of Portland. The anticline pitches 6° E. near Redlands, two miles north of Weymouth.

This simple pattern is disturbed in the axial region by many east–west strike faults. Dip faults are rare except in the region of the southern outcrop of Cornbrash. The strike of the main faults would, if theoretically projected, pass eastward to the Osmington and White Nothe area where both Intra-Cretaceous and Tertiary strike faults and folds are well displayed. The relation of the structures in the Weymouth district to those which may be dated farther to the east is one of the problems of the district.

The great Isle of Wight–Isle of Purbeck monocline, with an amplitude of about 2500 feet in the Lulworth district, passes westward into the sea south of Bat's Head (Fig. 1) on the east side of Weymouth Bay. On the west side of Weymouth Bay, along the projected strike of the vertical or overturned middle limb of the monocline, Jurassic rocks dip gently south, as already described, as if undisturbed by it. An explanation of the disappearance of this monocline, one of the greatest in the Mesozoic rocks of Britain, is one of the more important structural problems of the area.

4. STRUCTURAL DESCRIPTION

(a) The Faults of the Weymouth Anticline Core

Since the surveys of Bristowe and Strahan it has been known that the core of the Weymouth Anticline was cut by east–west strike faults. Detailed mapping has shown that these are more numerous than has been supposed. The course of the faults, and the names which are here given them, are shown on the accompanying maps (Figs. 1 and 2). The evidence for them will be briefly discussed, commencing accounts from the west, and proceeding from north to south.

(*i*) *The Chesters Hill Fault.* This was mapped by Strahan and Bristowe. At Berry Knap Cornbrash dips 15° N. towards the base of Chesters Hill which is capped by Cornbrash. Blue clay and septaria of the lower Oxford Clay crop out up to 100 yards NW. of Berry Knap, but then the shore becomes littered with loose Forest Marble material and some rubble of the

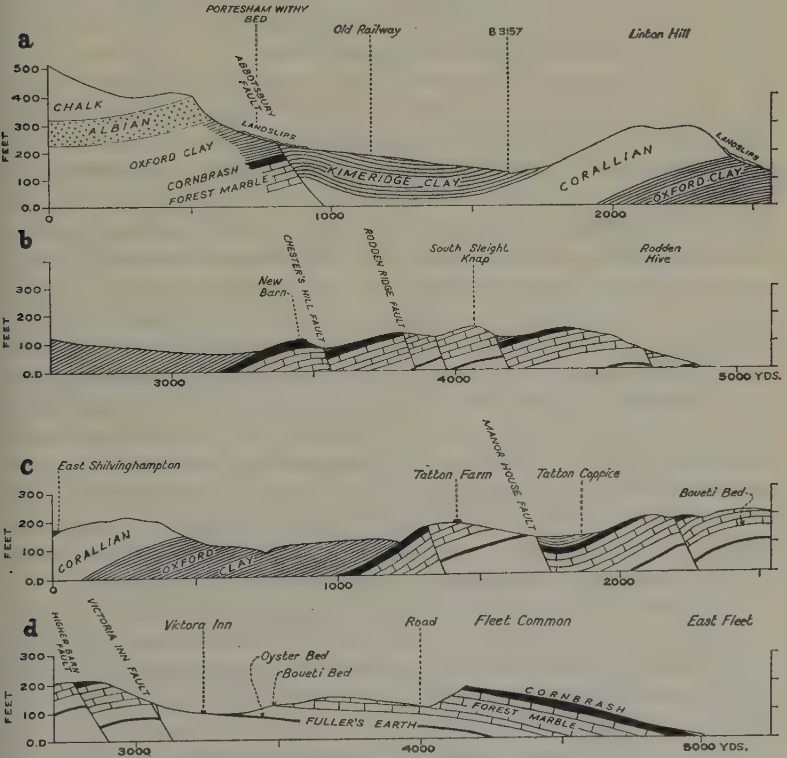

Fig. 3. Geological sections from north to south across the Weymouth Anticline. a, b. Section from the Abbotsbury Fault, one mile east-north-east of Abbotsbury, south-south-east to Rodden Hive on the Fleet shore. c, d. Section from the Corallian ridge at East Shilvinghampton south to the Fleet shore. The course of these sections is marked on an accompanying map (Fig. 1)

Cornbrash. No fossils of the *boueti* Bed have been found here, but the fault must have a throw of at least 150 feet south.

Eastward the Cornbrash ridges on either side of the fault converge and in Wyke Wood the Upper Cornbrash lies against Upper Cornbrash and the throw of the fault is very slight. During the past war a stick of bombs was dropped along the south-east edge of the wood, all within 150 yards. The most southerly crater, seventy yards NE. of an old Forest Marble quarry, showed upper Forest Marble rubble and fragments of Lower Cornbrash. The next crater, thirty-five yards NE. by N., exposed Cornbrash and from it was obtained a faunal suite of the Intermedia and Obovata Zones. The third crater, fifty-two yards to the NNW., exposed Oxford Clay, as did the last crater seventy yards farther to the north. The nearness of the Lower Cornbrash to the Oxford Clay is due to a fault downthrowing north which may be a continuation of the Chesters Hill fault, with a reversed throw, or a new strike fault.

(*ii*) *The Nottington or Nottington Village Fault*. The fault seen to commence at Wyke Wood can be traced for three miles to the east to the alluvial plain of the Wey north of Nottington Village after which the fault may be named. A subsidiary fault, the Rodden Ridge Fault, may be traced in a large field south-west of Wyke Wood, and it passes eastward to Rodden Ridge where for 1000 feet east of the road over the ridge it brings a reed-covered sliver of Oxford Clay on Cornbrash against Middle Forest Marble flags, a throw of forty or fifty feet. The fault then joins the Nottington Fault which has a throw of about fifty feet north, 800 yards NW., of Tatton Farm where Upper Cornbrash with *Macrocephalites* abuts Forest Marble flags. Beyond Hyde Coppice, Strahan mapped the Nottington Fault as swinging to the ENE. towards Holwell, but trenches for land-drains dug in a field 2000 feet N. of Buckland Ripers Church showed Oxford Clay covered by a thin soil with slipped chert gravel right up to the main road. The actual course of the fault, which lies farther south, was traced by auger. Where the Nottington Fault disappears beneath the alluvium of the Wey a minor fault joins it from the direction of Buckland Rectory which is responsible for two small outliers of Cornbrash.

Evidence of the continuation of the Nottington Fault has not been found in the wide expanse of Oxford Clay to the east. However, one mile due east of Nottington the Corallian ridge around Coffin Plantation (Fig. 4) is cut by two faults of southern downthrow which displace the Nothe Grit, join, and pass eastward.

(*iii*) *The South Sleight, Manor House and Nottington House Faults*. Parallel with the Rodden Ridge and Nottington Faults, but 300 to 400 yards to the south, is an important strike fault which downthrows to the south. In the west, near South Sleight Coppice, an outlier of Oxford Clay is brought against Lower Forest Marble, a throw of about 100 feet south, and there are minor faults associated with it. The main fault here, the South Sleight Fault, may pass east to join the Manor House Fault, well seen around the Langton Herring Manor House, which continues east to the River Wey. The wide expanse of Fuller's Earth between South Sleight Knap and Tatton Farm was missed in earlier surveys but both the *boueti* Bed and the Oyster Beds can be traced, although in places the latter seem to thin out.

Eastward from the Manor House a syncline in the lower Oxford Clay at Whitmore Coppice best explains the small outcrop of Upper Cornbrash which abuts sandy orange clay of the Kellaways Beds along the southern edge of Whitmore Plantation. Farther to the east the Manor House Fault may be traced through the hamlet of Buckland Ripers to the River Wey south of Nottington where the throw is seventy feet to the south.

On the east side of the River Wey, on the continuation of this fault, the Cornbrash ridge is cut by an east–west fault which has a northerly throw of thirty feet. This anomaly misled Strahan when mapping the Cornbrash hereabouts. The explanation is that another fault from the south-west, which downthrows more powerfully to the north than the Manor House Fault does to the south, joins the Manor House fault under the alluvium of the Wey, giving a net throw to the north. This new fault, the Nottington House Fault, is responsible for cutting out the Lower Cornbrash on the west side of the river and it brings the *boueti* Bed to within 100 yards of Upper Cornbrash, 350 yards N. of Harbour Bridge: the throw is about 100 feet to the north.

(*iv*) *The Coverwell and Higher Barn Faults*. Strahan mapped an unfaulted outcrop of Forest Marble 800 yards in width between Lower Farm and the northern margin of Wans Plantation yet remarked on the steps seen in the slopes down to the Fleet shore just to the west (Strahan, 1898, 7). The steps are caused by two strike faults which displace the flags of the Middle Forest Marble. The northernmost of these faults apparently passes westward to the Fleet shore 700 feet NW. of Rodden Hive Point where the Fuller's Earth Oyster Beds are brought against the clays which lie below it.

Farther east, to the south-east of Langton Cross, are two parallel faults which are thought to be the continuation of those just described. The northern fault can be traced almost to Colverwell Coppice, after which it is named. The southern fault, called the Higher Barn Fault, is responsible for two narrow outliers of Cornbrash. The relation of this fault to the Nottington House Fault, which downthrows in the opposite direction, cannot be traced with certainty in the Forest Marble outcrop south-east of Buckland Ripers.

(*v*) *The Faults around Pounds and Undercross Plantations*. Above Clay Hard Point, and just below the fifty foot contour near Pounds Plantation, the Fuller's Earth Oyster Beds are brought against the Lower Forest

Marble by a small strike fault which has not been traced eastward. The southern boundary of the outcrop of the Oyster Beds in the Plantation is also thought to be faulted since the Oyster Beds occur at a much lower altitude than would be expected if their northward rise from below the Coastguard Station continued.

Along the east side of Undercross Plantation, and almost parallel with the track leading from Langton Herring to Herbury, is the only important dip fault in the northern limb of the Weymouth Anticline. It can be traced for only 800 feet and appears to die out to the north: it may cross the isthmus of Herbury peninsula to the south.

(*vi*) *The Victoria Inn Fault*. This fault, which has a maximum downthrow north of about fifty feet, passes from the Fleet shore north-east of Herbury, past Bagwell Barn Cottages, and trace of it is lost on the hill slopes east-north-east of Chickerell Moor farm buildings. South of Longmoor Plantation is a curious triangular outcrop of Forest Marble which is limited to the north-west by a fault of southern downthrow.

(*vii*) *The Radipole Fault*. It is convenient to describe this fault from east to west. Strahan mapped the Radipole Fault in the east where it can first be traced in a field just east of Mountpleasant Farm, on the east side of the main Weymouth–Dorchester railway line. In this field, Oxford Clay on the north side of the fault rests against upthrown Cornbrash to the south, an estimated throw of 120 feet. The upthrown Cornbrash has been formerly worked from shallow pits. Farther west the fault has been traced by augering as far as Corfe Hill, where Cornbrash first appears on the northern side of the fault. West of the Wey Valley the *boueti* Bed crops out in the lower slopes of a large field 300 yards west of Causeway Farm, and along the hedge on the north side of the field *Ostrea hebridica* is common in the soil, showing that the Oyster Beds lie below. Forest Marble lies on the northern side of the fault here, and there are old quarries in the flaggy series in adjacent fields. Farther to the west the Forest Marble ridges converge until Forest Marble lies on both sides of the fault on Cold Harbour ridge. The fault continues to the west, however, and the *boueti* Bed is displaced about thirty feet north on the west side of the ridge.

(*viii*) *The Dip Faults of the Southern Cornbrash Ridge*. The main change to Strahan's mapping hereabouts is the complete elimination of the fault alleged to pass from near the Chickerell Church to the Fleet shore near East Fleet. Strahan considered that his fault cut out the Kellaways Beds on the Fleet shore, but Arkell (1948) has shown that all the zones of the Lower Oxford Clay and Kellaways Beds are present there. The new mapping has indicated the presence of several dip faults which divide the Cornbrash outcrop. The most westerly fault, which crosses Fleet Common, introduces a small triangular patch of Oxford Clay near the Fleet shore (House, 1958, 17, fig. 9).

(b) The Broadwey–Hamcliff–Ringstead Anticline

Within the broad arch of the Weymouth Anticline is a smaller and tighter fold which Strahan distinguished as the Broadway Anticline (now Broadwey: Strahan, 1898, 200). Its presence is indicated on the southern limb by a decrease in dip away from the axis of the Weymouth Anticline. The crest of the Broadwey Anticline is coincident with that of the main fold in the west, and it can hardly be distinguished there. By South Buckland and Redlands, however, the axis of the Broadwey Anticline has swung towards the east-north-east, whilst the structural crest of the main arch, as has been shown by Arkell (op. cit., 245), passes to the east-south-east.

Evidence that the Broadwey fold continues in the ridge formed by the Corallian Beds between Littlemoor and Hamcliff has been provided by the new mapping. The Corallian Beds in the region just west of Littlemoor Estate (Fig. 4) dip 9–10° NNE. The dip of the northern limb of the fold increases to the east so that it is 40° N. both at 300 yards north of the Riviera Hotel (Arkell *et al.*, 1947, pl. 19) and also near Eweleaze Barn, 650 yards NE. of Redcliff Point. The southern limb of the fold dips less steeply than the northern limb. Around South Down Dairy and Horse Lynch Plantation (Fig. 4) the Nothe Grit dips gently south. The fold axis can be traced through South Down Cottages and through the col on Jordan Hill and on to Hamcliff, where the anticlinal core is exposed in the cliff: here the Nothe Grit dips 15° S. on the southern limb 300 yards north-west of Redcliff Point. There is evidence suggestive of a synclinal flexure just south of the anticlinal axis in the slight northerly dips in the Nothe Grit exposed in the cliffs between Bowleaze Cove and Redcliff Point, but these may also be explained by 'bulging' due to the weight of the rest of the Corallian Beds above and the plasticity of the Oxford Clay beneath. At Redcliff Point the Oxford Clay dips 6° S. and at the southern end of Furzy Cliff it dips 2° S.: these dips are considered to be indicative of the southern limb of the Broadwey–Hamcliff Anticline. At the southern end of Furzy Cliff, two faults cut the Oxford Clay. The more southerly of these has a downthrow to the north and the calcite-veined and slickensided fault plane hades 30° S. This fault may well be the continuation of the Redcliff Point Fault.

(c) Notes on the Abbotsbury/Ridgeway Fault Belt

The fault belt which limits the Weymouth region to the north has been described in detail in the *Memoir* (Arkell *et al.*, 1947, 249 *et seq.*) and further details have been given in the *Bridport Memoir* (Wilson *et al.*, 1958, 113–29, 188). Some observations made whilst mapping this belt seem worthy of record: details will be given starting from the west.

At Portisham's Withy Bed, Strahan mapped downthrown Greensand resting on Kimeridge Clay on the north side of the Ridgeway Fault, and this interpretation is currently accepted (Wilson, op. cit., 189). The

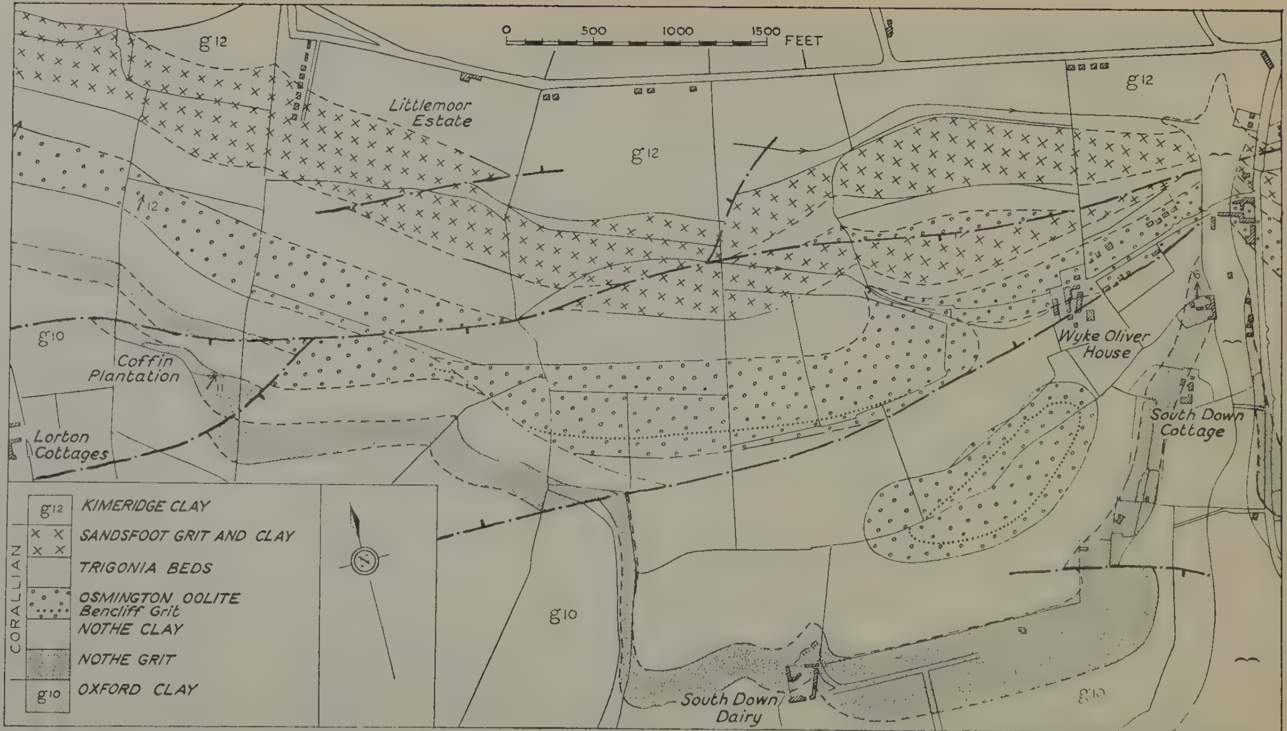

Fig. 4. The geology of the South Down Dairy Ridge, two miles north-north-east of Weymouth. Based on a twenty-five-inch survey by the author. The Bencliff Grit, here about ten feet thick, underlies the Osmington Oolite, but in places it has been mapped with it, and the same symbol is given on the map. The Ringstead Waxy Clays have been mapped with the Sandsfoot Grit

relations are much obscured by landslips hereabouts, but augering in one place, on the east side of the Withy Bed, proved a light blue and ochrous soft clay more typical of the Oxford Clay than the Kimeridge Clay, which is a stiff blue or grey clay. If the Gault does rest upon Oxford Clay here the tectonic picture is much simplified since the Tertiary Ridgeway Fault, downthrowing north, would pass directly into the Intra-Cretaceous Abbotsbury Fault exactly as it is known to do at depth farther to the east (this interpretation is shown on Figs. 1 and 3).

Some notes will be given to complement maps recently published of the Portisham and Friar Waddon areas (Wilson, op. cit., figs. 21 and 22). In the fields north-east of The Buildings, 1000 yards ENE. of Portisham Church, the Ridgeway Fault may be traced beneath a thin cover of flint and chert debris. Fragments of Portlandian limestone, with occasional ammonites, which have been found here suggest either that the subsidiary fault from Portisham continues parallel with the Ridgeway Fault for a longer distance than has been supposed or that a synclinal flexure is responsible for the exposure of younger beds.

In the Friar Waddon area evidence of the Cherty Freshwater Bed was found sixty-five yards south-west of the Well House, and in a field 260 yards south of Corton Barn, and was traced as far as the east side of Hell Bottom Quarry (330 yards west of the Bench) but was obscured for 340 yards south-east of the Bench. The outcrop is therefore more considerable than has been shown (op. cit., fig. 22). Blocks and fragments of Upper Greensand were noted in a field 300 yards NW. of Friar Waddon near the trace of the Ridgeway Fault. Presumably they came from a slice of Albian caught up in the fault.

In September 1959 a trench for a water pipe-line was dug from the reservoir one half-mile NNE. of St. Lawrence's Church, Upwey, southward through Gould's Bottom. The course of the trench is shown on an accompanying map (Fig. 5). The geology of Gould's Bottom was described by Arkell in some detail (1936, 101–3; 1947, 251) and details from his accounts are added to the map. The trench exposed a section across the Ridgeway Fault. A long section of the Purbeck Beds was exposed in the southern part of the trench and then for twenty-five yards Portland Stone, with *Chlamys lamellosa*, was seen which, at its conformable junction with tufaceous limestone at the base of the Purbeck Beds, dipped 18° S. The northern part of the Portland Stone exposure showed smashed limestone and chert passing towards a well-marked fault plane, hading 32° S., with iron-stained blocks of indurated limestone along it. Green, grey and red clays and silts of the Wealden and possibly Upper Purbeck Beds followed for twenty yards north of the fault and the dip was still southerly, though indistinct. Then another dislocation, not well defined, introduced a green glauconitic sand of Albian type for a further five yards, after which a brown

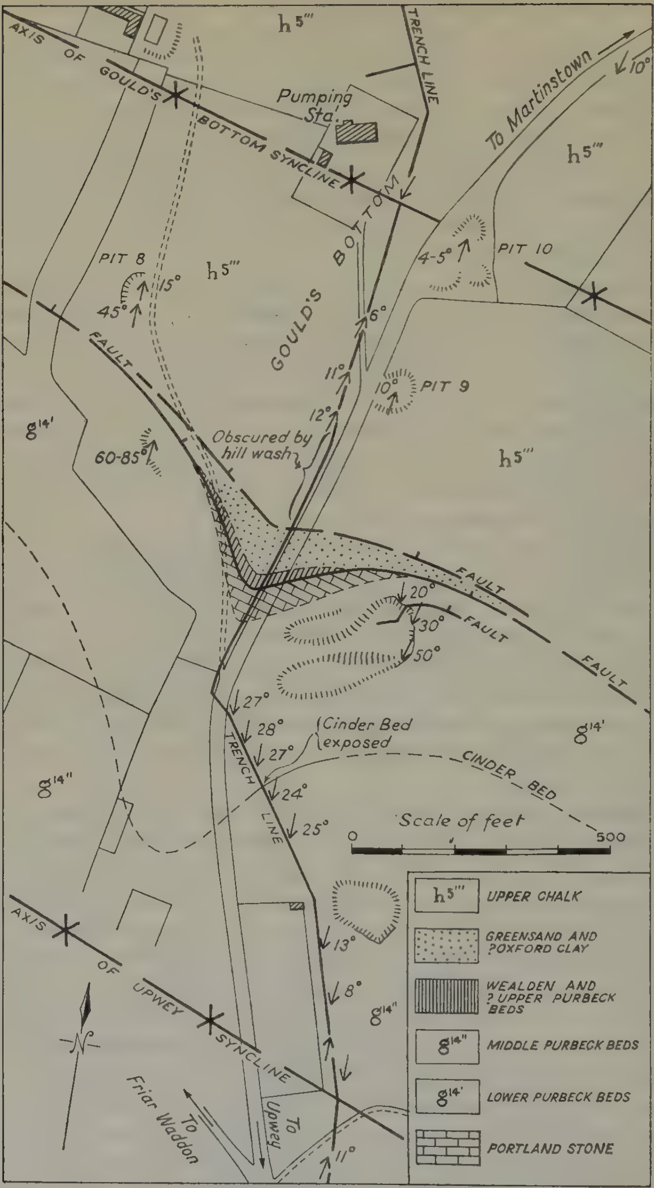

Fig. 5. The geology of Gould's Bottom, one half-mile north-west of Upwey, which is four miles north-north-west of Weymouth, showing the geology along a trench dug in September 1959. Pit numbers and some dip measurements are taken from the *Memoir*. For details see text

khaki clay was seen, apparently overlying the green sand at its first appearance. The clay was exposed for twenty-five yards until completely obscured by a hill wash of flint and chalk debris.

The trench exposures were surprising in that beds other than the Chalk and Purbeck have not been suspected in the valley heretofore. The brown clay was taken to be Oxford Clay, but no fossils or red nodules were found, and the clay was tougher than typical Oxford Clay. No fossils were found in the green sand.

A visit was made to the Bincombe railway cutting at Easter 1954 in the company of the late Dr. W. J. Arkell. A large block of ferruginous sandstone with Jurassic oysters, typical of the Nothe Grit, was found within twenty feet of the southward hading wall of brecciated chalk at the northern end of the fault inlier (Mottram, 1949, 182). The adjacent Oxford Clay here has recently yielded more ammonites from red nodules (House, 1955).

Special pits were dug in June 1954 at the Bincombe Inlier (Fig. 1) in an attempt to prove the presence of Forest Marble in the south-western part. Kimeridge Clay was found to rest against Cornbrash. Forest Marble is probably present in the western part of the inlier for the reasons given by Arkell (1947, 262).

The recent reinterpretation of the Poxwell area (Mottram & House, 1956) is important for a simple interpretation of the structures found in the boreholes at Chaldon Herring, farther to the east. It has been shown that the Poxwell Fault does not cut out all the Albian between Pixon Barn and Holworth, as shown on current Geological Survey maps (also Arkell *et al.*, 1947, pl. 19), but that it passes east into the Chalk at Moigns Down Barn. A continuation of the fault to the east gives an explanation for the anomalous dips and crush belt noted by Strahan north of Lord's Barrow (1898, 222; Arkell, op. cit., 277; Mottram, 1949, 177) and also provides a surface trace for the fault found at depth in the Chaldon Herring area (Figs. 1 and 6).

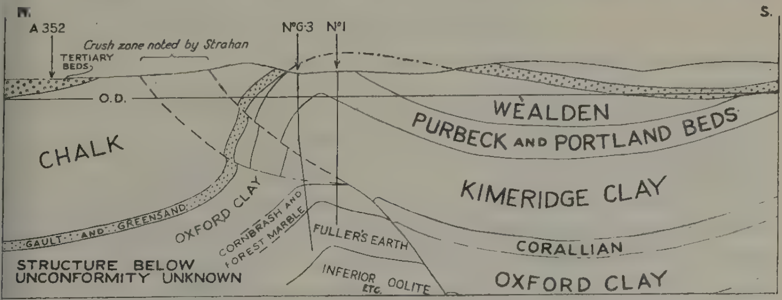

Fig. 6. Geological section north–south across the Chaldon Pericline, seven miles north-east by east of Weymouth, showing the boreholes of the British Petroleum Company. The syncline south of the pericline is the eastern extension of the Upton Syncline (Fig. 1). Length of section is 3050 yards. Vertical and horizontal scales the same

(d) The Chaldon Herring District

The Chaldon Pericline (Fig. 1) has recently been examined by the British Petroleum Company and three boreholes have been sunk to determine the structure. The position of the deep wells is shown on the accompanying map and section (Fig. 5). The details of the boreholes which are given below are summarised from unpublished well records and correlations made by A. E. H. Pedder, J. Harrison and G. J. Demaison.

(*i*) *Details of the Boreholes*

Chaldon Herring No. 1 Well. Summary of records by A. E. H. Pedder. Location, 02° 18′ 20.4″ West, 50° 39′ 15.3″ North.

	Depth in Feet	Apparent Thickness
Purbeck Beds	0–129	129
Portland Beds	129–351	222
Kimeridge Clay	351–1300	949
Lowest Kimeridge Clay, Corallian Beds and Oxford Clay absent		
Cornbrash (this and lower levels correlated with G–3 by electric log)	1300–?1330	?30
Forest Marble (dip at 1395 feet was *c.* 40°)	?1330–?1550	?220
Fuller's Earth (lowest 34 feet with 50° dip)	?1550– 1885	?335

Chaldon Herring No. G–3 Well. Summary of well records by G. J. Demaison. Location, 02° 18′ 22.4″ West, 50° 39′ 17.4″ North.

	Depth in Feet	Apparent Thickness
Purbeck Beds	0–185	185
Portland Beds	185–470	285
Kimeridge Clay (true thickness penetrated 365 feet, dip values increasing from 45° at 550 feet to 85° at 1000 feet)	470–1320	850
Lowest Kimeridge Clay, Corallian Beds and Oxford Clay absent		
Kellaways Beds (true thickness penetrated 56 feet, following gap of 59 feet)	1379–1440	61
Cornbrash (true thickness 24 feet)	1440–1468	28
Forest Marble (true thickness 175 feet, *boueti* Bed not identified at the base)	1468–1700	232
Fuller's Earth	1700–2244	544

Chaldon Herring No. G–2 Well. Summary of records by J. Harrison. Location, 02° 18′ 23.7″ West, 50° 39′ 3″ North.

	Depth in Feet
Head	0–10
Wealden Beds	10–540
Purbeck Beds	540–588

(*ii*) *Interpretation*

In two of these wells, No. 1 and G–3, a very important dislocation was encountered. No fault disturbance is known at surface in the area immediately surrounding the well sites apart from the crush zone noted by Strahan which has been already referred to and which is thought to mark a continuation of the Poxwell Fault.

In the light of the structures known at depth and at surface in the Poxwell Pericline it seems reasonable to infer that the fault found at depth at Chaldon is the eastward extension of the Poxwell Fault. This interpretation would mean that the Intra-Cretaceous Fault, throwing down to the south, must have been partially reactivated during the Tertiary movements as a reversed fault and, associated with the folding of the Tertiary pericline, the fault plane was folded just past the horizontal in the last stages. This interpretation is in agreement with structures known all along the Ridgeway and Abbotsbury Faults and, in particular, is the same hypothesis as Sir Edward Bailey (Arkell *et al.*, 1947, 268) offered to explain the Sutton Poyntz Inliers, although there the situation is rendered more complex by dip faulting. This interpretation is illustrated here by a diagram modified from a section prepared by Mr. G. J. Demaison after the writer had informed him of unpublished reinterpretations of the Poxwell structures (Mottram & House, 1956) (Fig. 5).

The possibility that a dip fault between boreholes No. 1 and G–3 might explain the apparent flattening of the fault plane at depth is thought to be unlikely since Tate was able to trace grit beds in the Wealden around the borehole sites (Arkell, op. cit., 276) and his map shows no evidence of a suitable displacement of the grit beds. There is, of course, no evidence of the course of the main fault below the position it was met in the wells. The steepening shown at depth on the section is based on the assumption that the folding of the fault plane is connected with the Tertiary folding of the pericline, and this is thought to be superficial. More sub-surface data is needed to confirm this.

5. THE DATE OF THE FOLDS AND FAULTS

Arkell has shown that the Ringstead Anticline is an Intra-Cretaceous structure, and there is evidence to show that this fold continues to the east at least as far as West Lulworth, where a borehole by the D'Arcy Exploration Company, started in 1939 and continued in 1947, proved Gault resting on Kimeridge Clay at −1580 feet O.D. (Falcon, *in* Arkell, 1951b, 178). To the west the Ringstead Anticline is known to link with the Hamcliff Anticline and it has been shown to continue into the Broadwey Anticline. There is thus support for the inference of Strahan (1898, 228–9) that the Broadwey Anticline is an Intra-Cretaceous structure.

This does not necessarily imply that the much broader arch, the Weymouth Anticline, is also of Intra-Cretaceous age. It is probable that the southern limb of the Weymouth Anticline is largely Tertiary in age by analogy with the Isles of Purbeck and Wight. If the northward dipping belts of Kimeridge Clay and Portland Beds are traced from Abbotsbury to the east-south-east they are seen to be cut across by southward dipping Gault and Greensand north of Black Head (Fig. 1 and Arkell, 1951a). This shows that the northern limb of the Weymouth Anticline is at least largely Intra-Cretaceous in age as Strahan believed (*supra cit.*).

The age of the faults in the core of the Weymouth Anticline can only be inferred indirectly since the overlying Cretaceous rocks have been eroded away. As a result of studying the tectonics of the Purbeck and Ridgeway Faults, Arkell suggested that 'the two east–west faults which run parallel for five miles from Buckland Ripers to the West Fleet, cutting only Jurassic rocks, may be of different ages' (1936, 115). Dated Intra-Cretaceous faults in the region, the Abbotsbury and Picnic Inn Faults (Arkell, 1951a, fig. 4) downthrow to the south and are normal. Dated Tertiary faults, such as the Ridgeway, Poxwell and Spring Bottom Faults, throw down to the north and are reversed. In the absence of other evidence it is reasonable to infer that the same rule may hold in the core of the Weymouth Anticline.

6. THE WESTWARD DISAPPEARANCE OF THE PURBECK MONOCLINE

The foresyncline of the Tertiary Isle of Purbeck Monocline passes westward into the sea at Middle Bottom, between White Nothe and Bat's Head (Fig. 1). Owing to coast erosion, aided by the easterly tilt of the monocline, evidence of this fold is lost in Weymouth Bay. The projected strike of the vertical or overturned Cretaceous rocks of the middle limb would pass across the bay into gently dipping Oxford Clay at Greenhill. The West Lulworth borehole has shown that the amplitude of the monocline must be about 2500 feet in the Lulworth district. Some explanation of the rather sudden westward disappearance of the monocline is required. Three hypotheses may be discussed.

(*i*) The Purbeck Monocline passes directly into the Weymouth Anticline. Arkell (*et al.*, 1947, 242) has suggested such a connection in terms of the crest-lines of the two folds. But genetically there is difficulty in connecting the middle limb of the monocline with the northern limb of the anticline, for the former is of Tertiary age and the latter, at least in the east, demonstrably Intra-Cretaceous.

(*ii*) The Purbeck Monocline may die out completely in Weymouth Bay. This theory was favoured by Strahan (1898, 220) who wrote: 'It [the monoclinal] must therefore die away rather abruptly in Weymouth Bay, not very far from the area where it attained its greatest intensity. In accordance with the general rule, simultaneously with the disappearance of this fold there commences the similar fault and anticline of Chaldon and Ridgeway, together with the intervening disturbance at Ringstead.'

(*iii*) The Radipole Fault may be the westward extension of a fault replacing the middle limb. The Radipole Fault diminishes westward at the rate of sixty feet per mile where it can be mapped. If the fault increases at the same rate to the east it could be a fault of some magnitude at White Nothe and there replace the middle limb of the monocline. In this way the apparent disappearance of the monocline is readily explained, and the crest-line continues in the Weymouth lowland on the upthrown side of the Radipole Fault.

Other workers are engaged on an investigation of the geology of the floor of Weymouth Bay, and the British Petroleum Co. Ltd. has a programme of boreholes in the area. It can be expected that light will be shed on several of the problems posed here when their results are published and available.

ACKNOWLEDGMENTS

The author must express his great indebtedness to the late Dr. W. J. Arkell both for the stimulus provided by his published work and for the personal encouragement and advice he gave during the course of the work: he also made valuable comments on the initial typescript. Thanks are also due to Mr. B. H. Mottram for a stimulating correspondence on the structures around White Nothe; both he and Dr. J. M. Hancock have kindly read and commented upon the typescript. The Durham Colleges in the University of Durham generously bore the cost of preparation of the text figures.

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Since this account was written, Falcon & Kent (1960, *Geol. Soc. Lond., mem.* 2, 49) have published details of the borehole drilled at Radipole, 1000 feet W. by N. of Causeway Farm (Fig. 1). They record a thickness for the Forest Marble quite at variance with the local thicknesses deduced both by the writer and by Sylvester-Bradley (1957, *Proc. geol. Soc. Lond.*, No. 1556, 26). There is a simple tectonic explanation of this anomaly. As is clear from the position of the well site (Fig. 1) and from the well records (kindly supplied by Mr. N. Falcon) the borehole first penetrated Fuller's Earth, and the records show shell bands with oysters between thirty-five and forty-eight feet. Below this clays with thin limestones followed to a depth of 163 feet: these were Fuller's Earth Clays. At some distance below this, which cannot be ascertained from the records, the reversed Radipole Fault will have been encountered, and the Forest Marble was then entered for the first time. If the *boueti* Bed was met at 242 feet, as Falcon & Kent suggest, then no more than at most seventy feet of Forest Marble can have been encountered. The anomalous thicknesses are therefore caused by a tectonic duplication of part of the succession by the action of the reversed Radipole Fault. The evidence suggests that this fault hades about 50° S.

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The Nature of the Millstone Grit–Carboniferous Limestone Junction of a Part of the North Crop of the South Wales Coalfield

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Received 29 October 1959

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ABSTRACT: Exposures of the junction between the Millstone Grit and Carboniferous Limestone along the eastern North Crop in South Wales are described. From their examination, it is suggested that the lowest Millstone Grit sediments were deposited in places prior to local induration of the calcareous sediments. Stratigraphical implications are discussed, together with theories of emplacement of the quartzitic sediments, and a sequence of events is proposed.

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1. INTRODUCTION

THE PRESENT STUDY is concerned with the relationship of the Millstone Grit to the Carboniferous Limestone between the Swansea and Clydach Valleys on the North Crop of the South Wales Coalfield (Fig. 1). Over this area, previous workers, notably Robertson (1927, 4), Robertson & George (1929, 38) and Evans & Jones (1929, 175), regarded the contact between these formations as unconformable. Their conclusions were primarily based upon the apparent indications of a systematic overstep of the limestone by the basal Millstone Grit accompanied by regular internal zonal overlap within the latter formation.

The overstep and overlap were believed to be a consequence of widespread uplift and folding of the limestones during late Visean and/or Namurian time. The uplift further resulted in the general southerly advance of a shoreline lying to the north of South Wales and in the emergence of a land area of considerable relief to the north, and also to the east—thus broadly located in the region of the present Usk anticline. Not until late Namurian time was the land area in the east finally submerged and buried beneath transgressive Millstone Grit sediments.

As most workers were mainly concerned with stratigraphy and zonal relationships, little detailed study was made of the nature of the actual contact between grit and limestone. At only two localities over a distance of thirty miles were features noted which suggested actual erosion of limestones prior to the deposition of overlying grits. The work of Owen (1954, 351) and Owen & Jones (1955, 457) showed that the overstep of the limestone was irregular between the Swansea and Neath Valleys. The scanty floral evidence further suggested that apparent zonal discrepancies in the basal Millstone Grit sediments could be ascribed partially to eastward thinning rather than entirely to regular overlap (Jones, 1958, 80).

In view of these modification of previous conceptions and the lack of a detailed description of the junction, the authors undertook its examination over a wider area. Much of the ground involved is composed of rough open moorland but the junction often stands out as a prominent northward facing scarp with Millstone Grit forming the summit. The boundary, easy to follow and map in the field, is rarely well exposed, being directly visible at only eight localities in the present area (Fig. 1).

2. GENERAL SUCCESSION

This is given in Fig. 2, the traditional faunal and lithological subdivisions of the Carboniferous Limestone being retained and no further subdivision has been attempted.

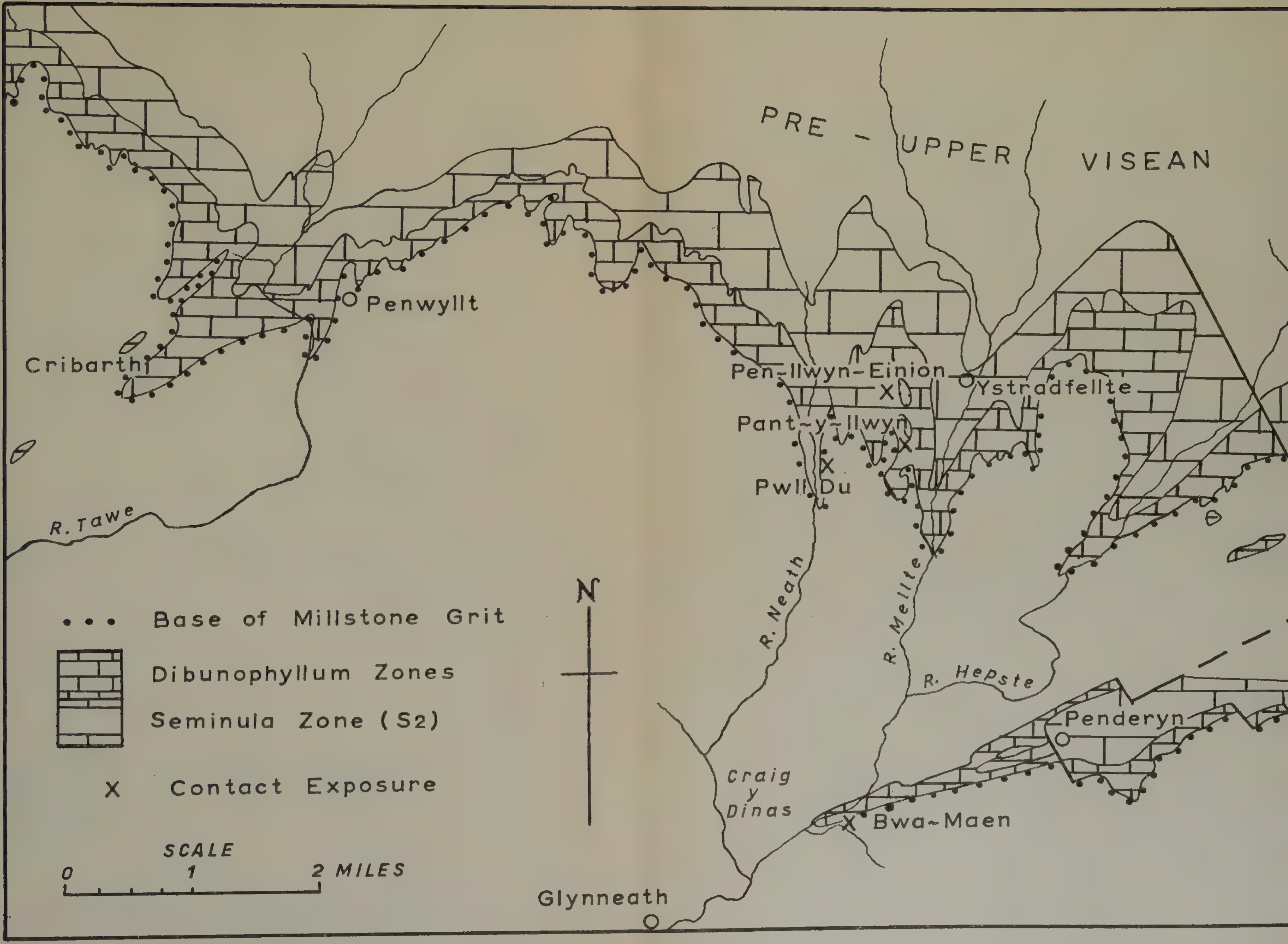
MAJOR LITHOLOGICAL DIVISIONS	STAGES	LITHOLOGICAL SUBDIVISIONS	ZONAL SUBDIVISIONS	LITHOLOGICAL CHARACTERISTICS AND THICKNESSES
MILLSTONE GRIT	NAMURIAN	BASAL GRITS	R_2 R_1 H? E	Mainly Sandstones with subordinate shale, clay bands and thin coals. Marine Shales towards top. Conspicuous bands of massive quartz conglomerate towards base. 80-450 ft.
CARBONIFEROUS LIMESTONE	UPPER VISEAN	UPPER LIMESTONE SHALES	"D_3"	Thin bands of calcite mudstone, black shales, secondary chert and fine detrital quartz. 0-30 ft.
		DARK LIMESTONE	"D_2"	Massive dark fragmental limestone with nodular chert, lenticular sandstones, shales, underclays and thin coals. 0-100 ft.
		LIGHT OOLITE	"D_1"	Coarse light to grey oolite with thin sandy developements especially at base. Thin coals, clays and concretionary beds towards top. Chert rare. 0-60 ft.
		DARK GREY LIMESTONES	"S_2"	Dark grey "flaggy" limestones and oolites. Extensively dolomitised at upper levels towards east. 200-350 ft.

Fig. 2. General succession within the area

3. DESCRIPTION OF THE CONTACT

(a) With the Upper Limestone Shales ('D_3')

The contact between the Millstone Grit and this subdivision is conformable near Llandybie, to the west of the Swansea Valley, and is believed


PRE - UPPER VISEAN
Penwyllt
Cribarth
Pen-llwyn-Einion
Ystradfellte
Pant-y-llwyn
Pwll Du
R. Tawe
R. Neath
R. Mellte
R. Hepste
Penderyn
Craig y Dinas
Bwa-Maen
Glynneath
N
• • • Base of Millstone Grit
Dibunophyllum Zones
Seminula Zone (S2)
X Contact Exposure
SCALE
0 1 2 MILES

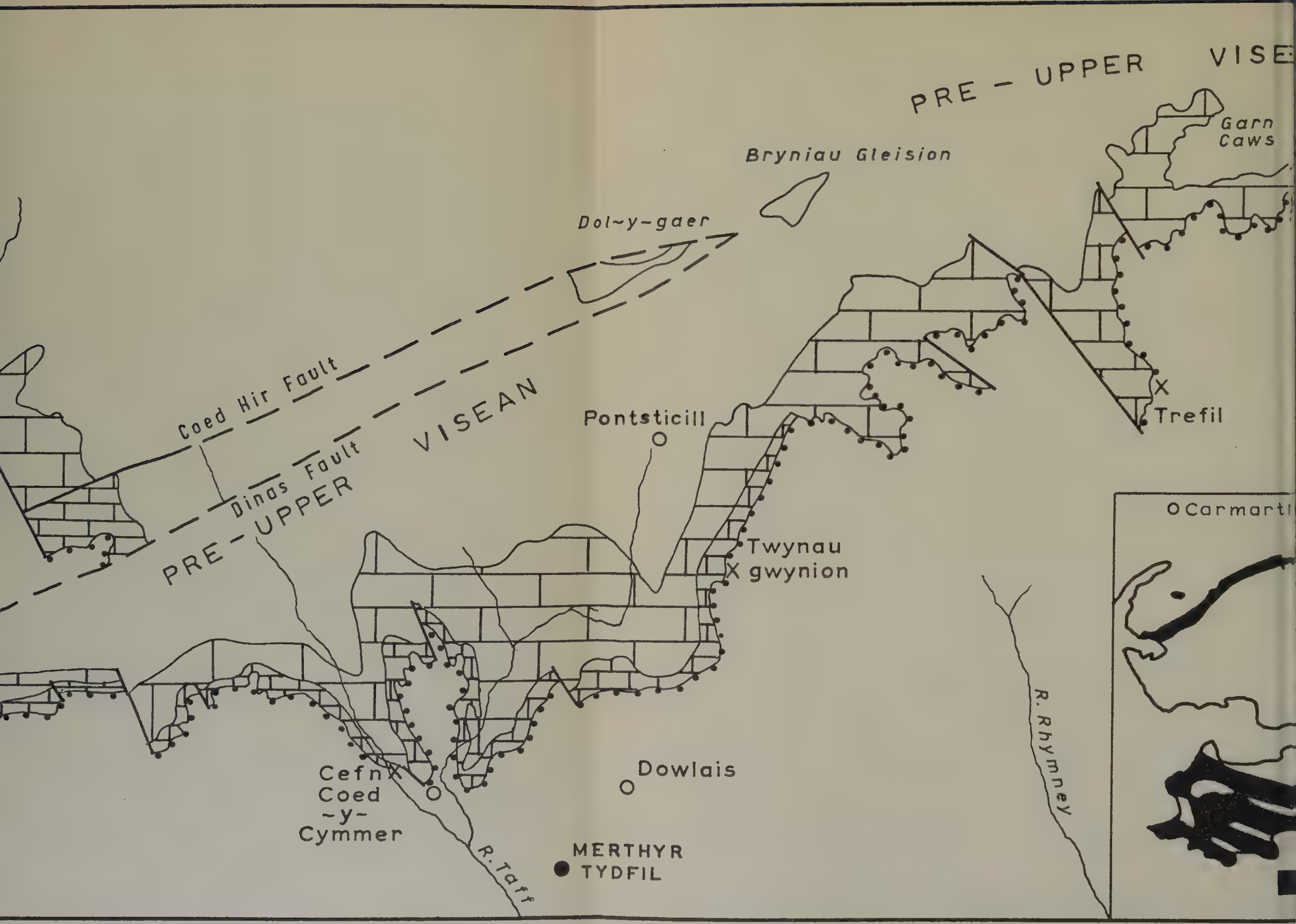
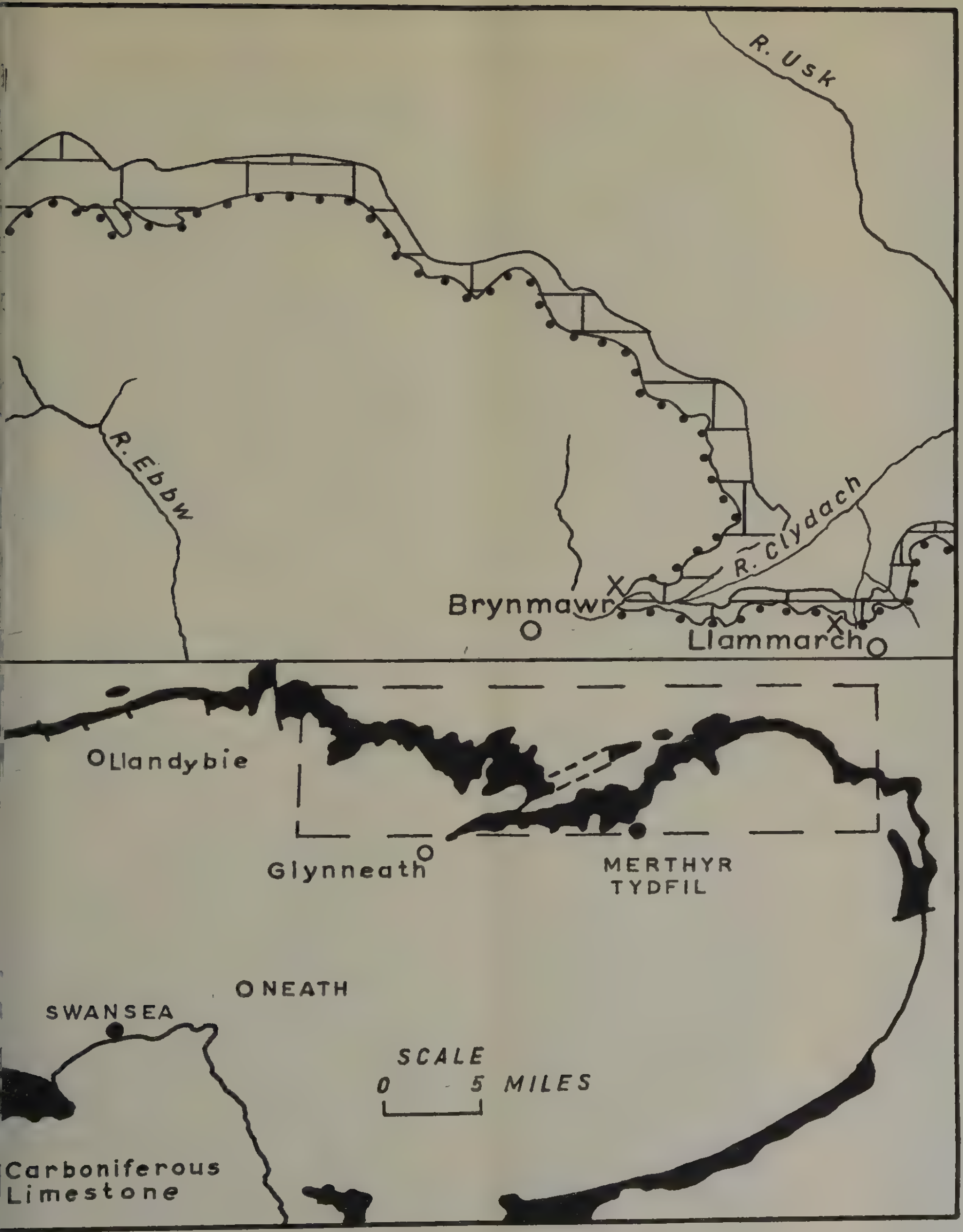


Fig. 1. Locality map for area studied


R. Usk
R. Ebbw
R. Clydach
Brynmawr
Llammarch
Llandybie
Glynneath
MERTHYR
TYDFIL
NEATH
SWANSEA
SCALE
0 5 MILES
Carboniferous
Limestone

[*To face p.* 240

to become disconformable eastwards (Ware, 1939, 201). No satisfactory exposures of the junction occur between the Swansea and Neath Valleys and this and the general absence of goniatites make it difficult to assess the extent of any time-gap. On the South Crop of the Coalfield goniatites indicating a P_2-E_2 age have been recorded from boreholes in the Upper Limestone Shales (Anderson, 1956) and it is probable that the North Crop sequence is of the same age. If this is correct and combined with the age of the flora at the base of the Millstone Grit (Dix, 1933, 163; Ware, 1939, 201; Jones, 1958, 80) then any interval would not be great in zonal terms. However, it should be noted that the North Crop sequence is thin and appears to be absent in places between the Swansea and Neath Valleys. The small exposure at Mellte Bridge (Owen & Jones, 1955, 457) shows that Upper Limestone Shales were deposited as far east as this point but their absence elsewhere between these valleys cannot be definitely ascribed to non-sequence or subsequent removal by erosion. It is significant that although the 'Black Lias' type facies is present over wide areas the sequence occurring to the east of the coalfield, e.g. north of the Bristol district, is mainly arenaceous (George, 1958, 263; Smith, 1930, 333).

(b) With the Dark Limestones ('D_2')

These limestones are present beneath the Basal Grits from Craig-y-Dinas to the Cefn Cadlan area north-east of Penderyn, the contact being exposed at four localities, viz. Bŵa-Maen, Pant-y-llwyn, Pen-llwyn-Einion and Pwll-du.

(*i*) *Bŵa-Maen*. A good section of the contact is exposed in the bed of the River Sychryd near the silica levels for some 100 yards. There is no discordance of dip between grit and limestone but the actual junction is irregular with pipes and funnels of grit from the main overlying conglomerate bed penetrating the limestones to depths of twelve feet (Fig. 3). Along some of the pipe margins the contact between grit and limestone is sharp but with others the lithological boundaries are vague and transitional. Predominantly quartzitic masses thus merge through pebbly and gritty limestones to normal dark limestone in short distances—usually a matter of inches. Isolated 'trains' and stringers of pebbles also occur but have no regular alignment. The limestone itself is unaltered, and in primary lithology consists of a dark fragmental (chiefly crinoidal) sediment with considerable admixture of detrital quartz which is often so abundant as to make up thin sandstone layers. In marginal areas it fills interspaces between the quartz pebbles. Deformation of bedding in the limestone is noticeable near some of the pebbly and quartzitic masses (see Plate 7A). Organic fragments, chiefly fish teeth and crinoidal ossicles, occur abundantly in the lower quartzitic material, but only rarely in the basal layers of the overlying main conglomerate bed.

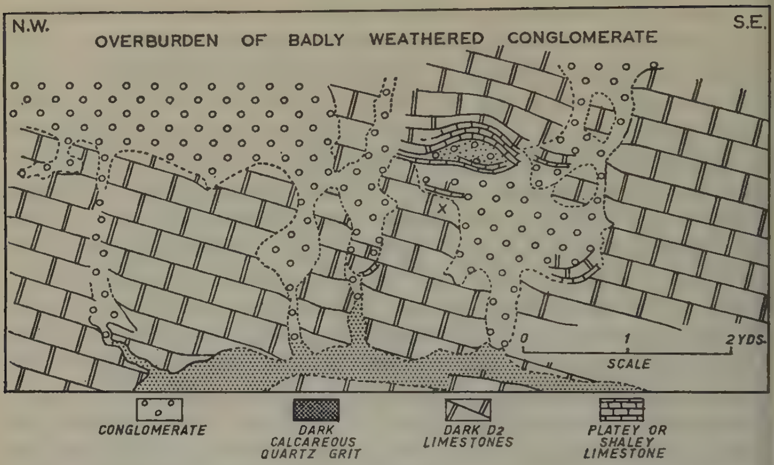
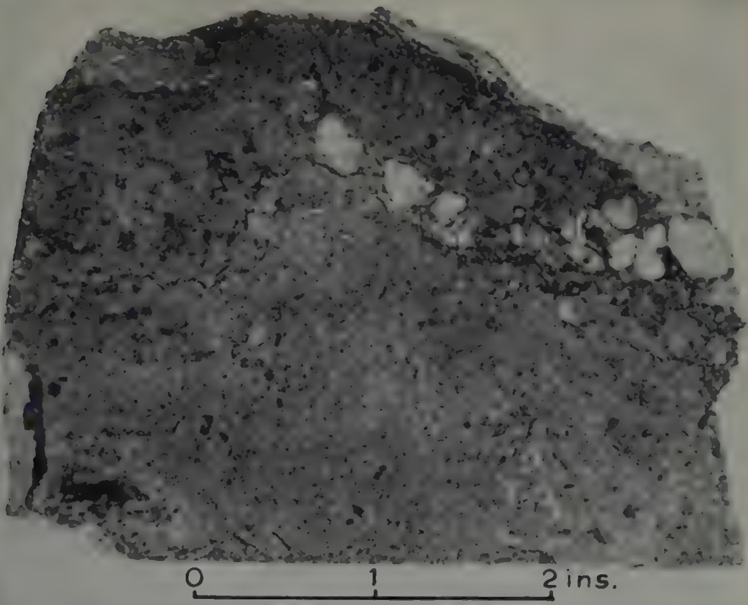

N.W.
S.E.
OVERBURDEN OF BADLY WEATHERED CONGLOMERATE
P
X
0
1
2 YDS.
SCALE
CONGLOMERATE
DARK CALCAREOUS QUARTZ GRIT
DARK D2 LIMESTONES
PLATEY OR SHALEY LIMESTONE

Fig. 3. Section of Millstone Grit–Carboniferous Limestone contact near Dinas silica levels

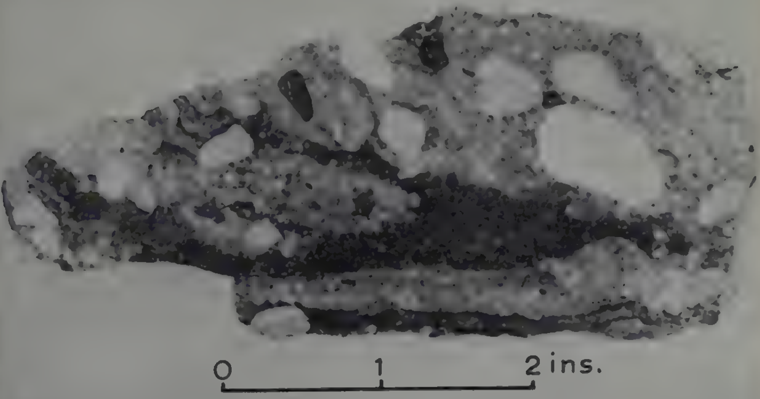
The irregularity of the junction, the sharp contact in places and abrupt lithological change would suggest disconformity, but the vague contacts and deformation of limestone bedding suggest some degree of plasticity in the limestone during deposition of the pebbly grits.

(*ii*) *Pant-y-llwyn.* A small cave and bluff exposes the contact 400 yards north of Pant-y-llwyn farm where eight feet of bedded sandstones and dark pebbly grits overlie ten feet of D_2 limestone. Although dark and crinoidal, the uppermost calcareous sediments show more diversity than at Bŵa-Maen. Clay with calcareous nodules occurs with irregularly developed areas of slightly dolomitised calcite mudstone, intercalated with the more typical pellet and fragmental crinoidal limestone. Detrital quartz, although still appreciable in amount and present in most of the calcareous lithologies, is not so abundant here as at Bŵa-Maen. The actual contact is, however, generally similar in its irregularity, with pipes of pebbly grit penetrating to depths of ten to twelve feet below the main bed of conglomerate. Some sagging of bedding can be seen in the pebbly grit infilling of some of the pipes. Contacts along some of the pipe margins are sharp but this can be ascribed in part to recent solution weathering leaving a small gap between pebbly grit and limestone. Similar small gaps occur along the contact with the main conglomerate bed. Some pipe margins and contacts, however, are transitional, so that the quartzitic detritus is intimately admixed with the pellet and crinoidal layers (see Plate 7B).

The marine origin of the clay patches is shown by the occurrence of Productids in the calcareous concretions and the presence of the small



A



B

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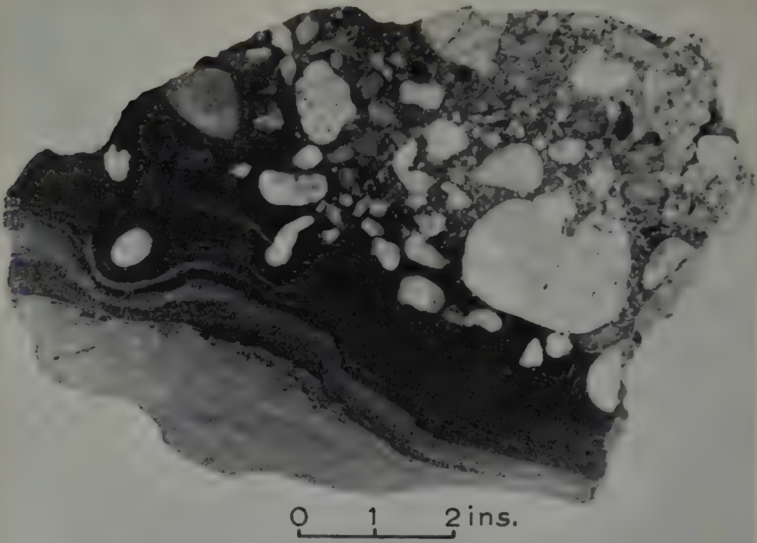
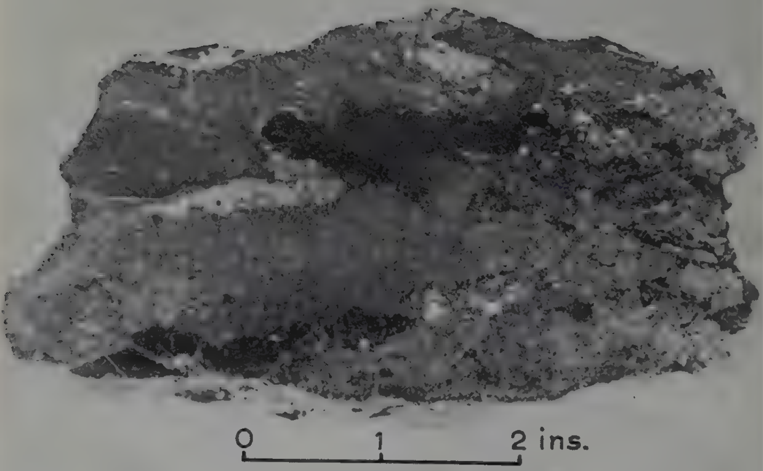


A



B

caves is due to the clays being washed out by the small stream at present flowing over the rock face.

(*iii*) *Pen-llwyn-Einion.* The contact exposure at this locality was described by Robertson (1933, 54), who concluded that it showed evidence of erosion of limestone prior to the deposition of the overlying grit. The actual exposure is some five yards long and the combined thickness of grit and limestone exposed is nine feet. The limestones are coarsely crinoidal and the contact features observed at other localities are present and are shown in Fig. 4.

(*iv*) *Pwll-du.* Occasional exposures of the junction occur in the deep ravine of the upper Neath and, where seen, all contacts are sharp and irregular. Tabular masses of chert occur in the lowermost two feet of conglomerate but the relationship is obscure. Recent weathering has emphasised the irregularity of the upper surface of the limestone and no merging or transitional areas are observable. The available evidence at this locality would suggest that the contact is disconformable but no pipes or funnels of grit occur.

(c) With the Light Oolite (‘D_1’)

The dark limestones of the D_2 subzone disappear east of Cefn Cadlan, their place beneath the Millstone Grit being taken by the massive, coarsely oolitic, beds of the D_1 subdivision. The disappearance is not regular or systematic and whether due to overstep or ‘feather edging’ is difficult to determine in view of the severe faulting associated with the Vale of Neath Disturbance. Owen (1954, 351) suggested that the thickness variations of the dark limestones in the area could be due to pre-Namurian arching along the Disturbance. East of Cefn Cadlan the junction with the Light Oolite is exposed only at two localities—Twynau gwynion and Cefn-Coed-y-Cymmer.

(*i*) *Twynau gwynion.* North of Dowlais the contact is to be seen in old shallow surface quarries and once again it is not straightforward. As with the D_2 limestones there are sharp and merging areas (Plate 8A) which result in a considerable variation in lithology, ranging from normal crinoidal oolite with abundant *Productus hemisphaericus* J. Sow., and *Lithostrotion martini* Edwards and Haime, through sandy and pebbly calcareous beds to coarse crinoidal sandstones. In places the calcareous sandstones have been decalcified and strongly impregnated with iron. These can also be seen unweathered and thus there is no doubt that the decalcification and iron enrichment is a recent leaching phenomenon and the resulting material is not residue derived from an ancient calcareous erosion surface.

(*ii*) *Cefn-Coed-y-Cymmer.* Forty feet of the contact is exposed as part of a cliff with small caves on the north bank of the River Taff near Merthyr Tydfil. The base of the conglomerate shows some undulation and for some

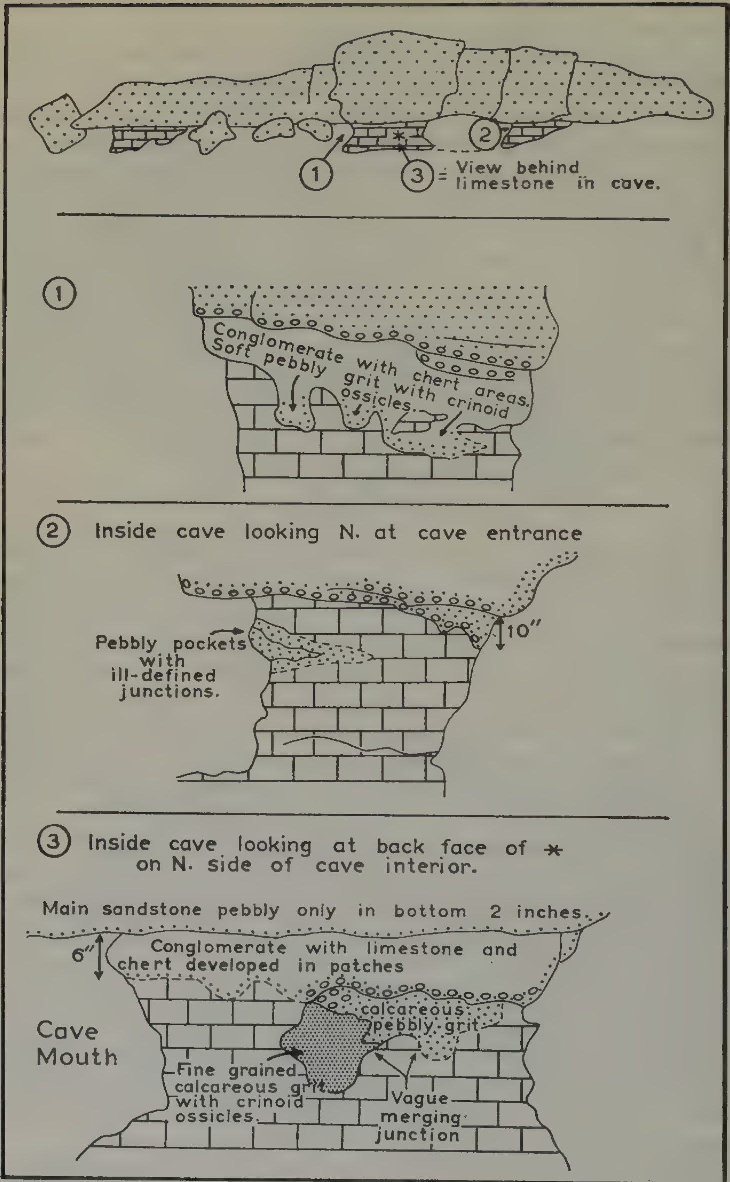

2
1
3 = View behind limestone in cave.
1
Conglomerate with chert areas.
Soft pebbly grit with crinoid ossicles.
2 Inside cave looking N. at cave entrance
Pebbly pockets with ill-defined junctions.
10"
3 Inside cave looking at back face of ✱ on N. side of cave interior.
Main sandstone pebbly only in bottom 2 inches.
6"
Conglomerate with limestone and chert developed in patches
calcareous pebbly grit
Cave Mouth
Fine grained calcareous grit with crinoid ossicles.
Vague merging junction

Fig. 4. Contact near Pen-llwyn-Einion

distance the lowest layer is a large tabular mass of chert lying parallel to the limestone bedding. The upper surface of the chert is 'welded' to the Basal Grits while the lowermost grit layers contain unworn crinoid ossicles with the central canal filled with quartzitic material identical with the matrix, indicating that they are indigenous and not derived. They also demonstrate the marine origin of the lowermost Millstone Grit sediments. However, apart from these fossils, nothing is present which would indicate anything other than a disconformable contact at this locality.

(d) With the Main Seminula Zone ('S_2')

Eastwards of Twynau gwynion the junction is entirely obscured by scree and drift for four miles and the exact manner of disappearance of the Light Oolite is uncertain. South-east of Ponsticill, S_2 limestones appear as the highest beds beneath the Basal Grits, but once more exposures are poor and the contact is visible only at two localities over a distance of nine miles, i.e. at Trefil and in the Clydach Valley.

(*i*) *Trefil*. The main exposures are here found in old quarries partially obscured by grass and scree; dolomitisation and silicification add further difficulties. The general subsequent nature of the dolomites, discussed by Robertson (1927, 39) and amply confirmed for the Tournaisian beds below by George (1954, 291), is well shown by the development of dolomite rhombs. These often increase rapidly in abundance until the normal dark limestone is completely replaced in a few feet, but rarely does original unaltered limestone occur beneath the Basal Grit. The grits themselves are strongly iron-stained for a few feet at their lower levels and have patchy developments of red and green clay. There is thus a contact zone rather than junction along a single surface. Quartzitic material with dolomitic cement often occurs below the main bed but the whole zone is badly weathered. The subsequent nature of the chert masses is indicated by the presence in them of *Seminula ficoides* and pectenoid lamellibranchs and other fossils. The masses appear to have caused disturbance in the overlying grits which arch over them, the arching being possibly due to expansion consequent on silicification rather than a 'supratenuous' depositional effect.

(*ii*) *The Clydach Valley* (*near Brynmawr*). Exposures of the contact occur along the sides of the gorge and in the tributary valley of Llammarch dingle, the latter being the most accessible. Dixon (1931, 161) considered the irregularities of the upper surface of the limestone as here showing evidence of erosion prior to deposition of the Millstone Grit. A purely erosional interpretation of the contact is, however, an over-simplification. In the gorge of the River Clydach, near the pumping station, the junction is very irregular but the anomalous position of the lower beds below marker horizons in the basal Millstone Grit suggest the possibility of subsidence and warping as the major cause of irregularity. In Llammarch dingle, as with Trefil quarries, a contact zone rather than abrupt lithological change characterises the junction and a section near the railway tunnel is given below:

	ft.	in.
Sandstone		
Shale, grey, coaly with plants	1	6
Sandstone, dark, thin bedded	5	0
Sandstone, light, pebbly coarsening to conglomerate with intermixed light grey chert masses, areas of green shale and mudstone, dolomitic cement in places, base markedly irregular ...	7	0
Dolomite, crystalline with patches of quartz sand	3	0
Dolomite, buff, crystalline	—	

4. THEORIES CONCERNING THE NATURE OF THE CONTACT

(a) Interpretation Favoured by the Authors

The contact of Basal Grits and dark limestone (D_2) at each locality has features indicating that it is not a straightforward plane of marine abrasion or unconformity, developed as a result of Namurian transgression. The most significant feature is the presence of quartzitic detritus well below the main bed of conglomerate accompanied by deformation in the surrounding limestone. The deformation around and near the pebbly quartzitic bodies is clearly not a result of post-lithification processes but must have occurred while the calcareous sediment was in a plastic state. This is shown by the intermixture of the quartzitic material and unaltered sandy limestone, contortion of laminae around and below pebbles and evidence of squeezing of calcareous material by pebbles. There is no evidence to suggest that large-scale open slumping or simple load casting were the major causes of these features. Small-scale load casting and flap-like ‘turnovers’ of thin laminae are present beneath the main conglomerate at Bŵa-Maen and Pant-y-llwyn but it is considered that the bulk of the contortion was produced as a result of differential stresses set up during compaction. These resulted in the pressing of quartz sand and pebbles into unindurated calcareous muds and, in places, movement of connate fluids which leached out calcareous material leaving an insoluble residue of quartz sand. This is to be seen in the presence of sandy ‘haloes’ around the pebbles and lenticular ‘stringers’ of quartz sand.

Diagenetic effects, such as re-sedimentation and formation and infilling of microcavities similar to those described by Sander (1951, 26) and Bathurst (1957, 33), are present both along and away from the contact with the D_2 limestones. They are better shown, however, in the D_1 limestones where the coarse oolite texture probably offered greater opportunity for post-depositional solution activity. While some admixture of quartzitic and calcareous material in the D_1 sediments can be ascribed to diagenesis, plastic deformation is present, and indicates that the Light Oolite was at most only partially indurated locally, when the basal Millstone Grit sediments were deposited.

Whether any prolonged emergence and desiccation of the upper calcareous surface occurred, prior to or during the introduction of the quartzitic material, is not possible to ascertain directly. There are no typical features, such as shrinkage cracks, clay dykes and curved polygons, to suggest exposure to the atmosphere, although this conceivably may be due to lithological unsuitability. The degree of coherence of the upper surface of the calcareous muds is thus problematical but the available evidence indicates that the introduction of quartzitic material was subaquatic and that it sank into muds which were sufficiently rigid to resist gross deep disturbance.

(b) Alternative Explanations

The pebbly masses of D_1 and D_2 age could be interpreted as infillings of relict sink holes formed during submergence and burial of a former pre-Namurian karst topography, the pipes and funnels with sharp contacts being thus explained as simple Neptunean dykes and the intimate admixture of quartzitic and calcareous material as incorporated weathered residuum. However, this explanation fails to account for the deformation near the pebbly areas, the general lack of limestone and chert boulders and absence of widespread silicification. There is also no typical material which could be considered as a weathered residuum off an old calcareous erosion surface. These observations also invalidate the possibility of post-Namurian sub-surface solution and cavern roof collapse as a primary mechanism similar to that described by Bretz (1940, 379). The contact also does not resemble the former 'open' collapsed surface developed on Carboniferous Limestone during Triassic times described by Dixon (1921, 162) in Pembrokeshire and that on the North Crop of Tertiary age described by Thomas (1954, 220). The contact features are on a much smaller scale and there is an absence of the spectacular brecciation noted elsewhere, in fact of any brecciation. Thus the deformation along the contact with the D Zone limestones cannot be ascribed to sagging due to collapse or subsidence of lithified limestones. In areas where the contact is with S_2 Zone limestones, subsequent dolomitisation makes it difficult to assess whether the quartzitic material below the main bed was emplaced in a manner similar to the D Zone limestones. Thus it could not be determined whether any of the transition zone material represents weathered remnants of thin original non-calcareous upper beds or concentrations of fine clastics originally present in the limestone or sub-surface solution residuum. In the absence of any conclusive evidence of reworking during later transgression (either marine or deltaic) and of undoubted sub-aerial weathering, it is suggested that the major cause of irregularity of the contact in S_2 Zone areas was intra-Namurian sub-surface solution. This probably accentuated the relief of a contact already much affected by adjustments due to earlier secondary dolomitisation.

5. STRATIGRAPHICAL IMPLICATIONS OF TIME OF EMPLACEMENT

The emplacement of pebbly grits in unlithified calcareous sediments necessitates a revision of the age of the lowest beds of the Millstone Grit facies. It is apparent that no great time-interval in zonal terms can separate the uppermost D_2 limestones and the lowermost quartz conglomerates locally, unless the calcareous sediments remained in an unlithified state for a prolonged period. The presence of *Lyginopteris stangeri*, *L. fragilis* and *Diplotmema adiantoides* twelve feet above the conglomerate at Bŵa-Maen and Penderyn (Jones, 1958, 77) suggests that they would have remained unindurated through a period spanning 'D_2' (P_1), 'D_3' (P_2) and part of the E Zones. As such a period without induration is unlikely for calcareous sediments it is therefore probable that some measure of contemporaneity exists, at least locally, between the quartz conglomerates and D_2 limestones, i.e. some of the Millstone Grit facies is of Upper Visean age. Such a view of age would imply a diachronous behaviour of the base of the Millstone Grit, in a direction, opposite to that suggested by Evans & Jones (1929, 174) and Pringle & George (1956). 'Younging' is now shown to occur westwards and is further supported by evidence of plasticity in the Light Oolite at Twynau gwynion.

6. SUGGESTED SEQUENCE OF EVENTS

The general tilting of the southern flanks of St. George's land operative in Upper Visean times (George, 1958, 262), intensified in *Dibunophyllum* Zone times and led to marine regression and southerly advance of the shoreline lying to the north. The advance of the shoreline was not uniform but proceeded more rapidly along headlands or shoals projecting from the shoreline. The trend of the headlands anticipated that of later important Armorican structures, such as the Neath Disturbance and Usk Anticline, and calcareous deposition ceased and quartzitic detritus was introduced earlier in the vicinity of these axes. The absence of various *Dibunophyllum* Sub-zones in the eastern portion is thus probably due as much to non-deposition as to apparent later Namurian transgression.

ACKNOWLEDGMENTS

It is a pleasure to acknowledge the authors' indebtedness to Professor T. N. George, Mr. W. D. Ware and other colleagues and friends for many helpful and fruitful discussions and to the technical staffs of King's College, London, and University College, Swansea, for the preparation of maps, figures and rock sections.

EXPLANATION OF PLATES

PLATE 7

A. Quartz pebbles embedded in oolitic and pellet limestone. Junction with Light Oolite (D_1), Twynau gwynion

B. Interfingering of Dark Limestones (D_2) and pebbly quartzitic material. Pant-y-llwyn

PLATE 8

A. Junction at Bŵa-Maen with depressed and contorted laminae in dark limestone. The thin light-coloured bands are sandy laminae within the Dark Limestone (D_2). Specimen is taken from point marked 'x' in Fig. 3

B. Intimate mixing of calcite mudstone, pellet and crinoidal limestone with pebbly quartzitic material along junction at Pant-y-llwyn

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A New British Devonian Alga, *Palaeoporella lummatonensis*, and the Brachiopod Evidence of the Age of the Lummaton Shell-Bed

by GRAHAM F. ELLIOTT

Received 3 October 1960

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ABSTRACT: A codiacid alga, *Palaeoporella lummatonensis* sp. nov., is described from the Lummaton shell-bed in the marine Devonian limestones of Torquay, Devonshire. The microfacies and fossils of this bed are considered to confirm earlier views as to the initial current-accumulation of the material. The age of the bed, from brachiopod and other evidence, is thought to be top Middle Devonian (Upper Givetian).

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1. INTRODUCTION

DURING EXAMINATION of thin-sections of Devonian limestones from the Torquay district remains of a small cylindrical alga with radiate structure, evidently a dasyclad or codiacid (chlorophyceae or green algae), were noted. Further preparations revealed this to be a species of *Palaeoporella*, a genus known from the Lower Palaeozoic. Devonian algae are uncommon when compared with those described from the Lower Palaeozoic or from the Carboniferous and Permian (cf. Johnson & Konishi, 1958): the age of the bed from which the alga comes is discussed on the evidence of the associated brachiopods.

2. SYSTEMATIC DESCRIPTION

Class: *CHLOROPHYCEAE*

Order: SIPHONALES

Family: CODIACEAE

Genus: **PALAEOPORELLA** Stolley, 1893

Palaeoporella lummatonensis sp. nov.

(Plates 9, 10)

Diagnosis. Calcified elongate-cylindrical units (segments) of circular cross-section, diameter up to 2.5 mm. (exceptionally greater), observed length 10 mm. incomplete; medullary core lightly calcified with indistinct longitudinal fibres, cortex showing irregular distally directed radiate fibres, often bifurcating, and usually widening sharply at their outer terminations.

Holotype. The specimen figured in Plate 9, Fig. 1, from the Lummaton Shell-bed, top Middle Devonian (Upper Givetian), small western quarry, Lummaton Hill, Torquay, Devonshire. British Museum (Natural History) Department of Palaeontology, registered number V.42603.

Paratypes. The specimens figured in Plate 9, Figs. 2, 5; Plate 10, Figs. 1, 2, 4; same locality and horizon, registered numbers V.42604, 42607–9 incl.

Other material. Numerous random sections in material from the same locality and horizon.

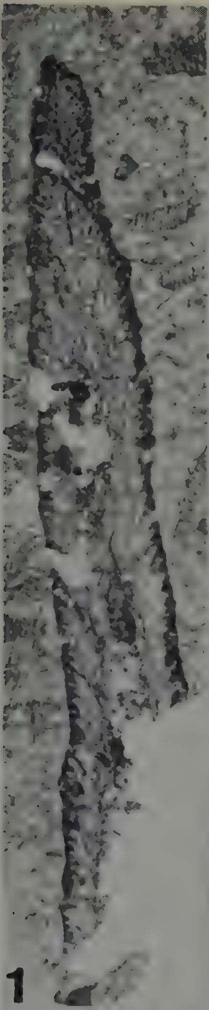
Description. The segments of this species are typically long and cylindrical, rounded in cross-section and with rounded-conical terminations. Exceptionally they are curved, and there is a minority of shorter stouter segments, as is to be expected by analogy with Recent codiacids. Segments which show bifurcation themselves have not been found; the single segments may quite probably have budded off in pairs from the terminations of early single segments, as is so often the case, but the material examined did not give any evidence on this point.

The medullary cores show indistinct and irregular thin longitudinal threads: presumably in life calcification was poor in this region. The boundary between core and cortex is usually indefinite or irregular. In the fossils these cores are usually replaced by coarse dolomite crystals, and the original structure obliterated.

The threads of the cortex are seen to individualise subcentrally, in well-preserved examples, as thin wavy lines: they run in gentle divergent curves distally and outwardly, then thicken and bifurcate in a characteristic 'blot and smear' pattern, and curve more sharply to terminate in abrupt triangular thickenings at the outer surface. These subdermal terminations are probably blurred preservation due to diagenesis of original bunches of numerous closely set fine terminal threads, by analogy with related living forms. Exceptionally this finer structure is indistinctly visible.

Discussion. Although some transverse cuts of small segments of this alga (e.g. Plate 9, Fig. 4) suggest a small dasyclad such as the Carboniferous *Anatolipora* (Konishi, 1956), the absence of a well-defined central tubular cavity and the irregular arrangement and nature of the algal threads, rules out reference to the Dasycladaceae and indicates the Codiaceae. Two main groups of calcified genera are referred to this family: the extinct nodular

PLATE 9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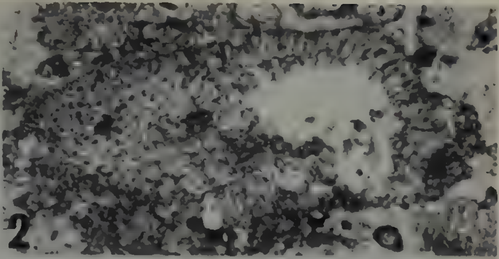
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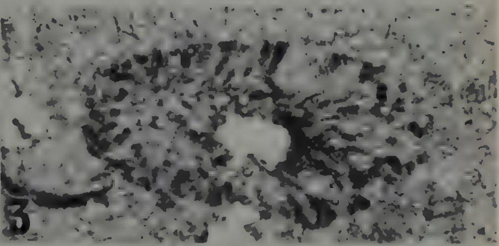
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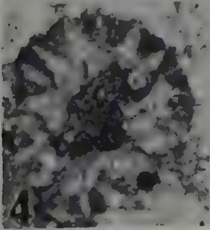
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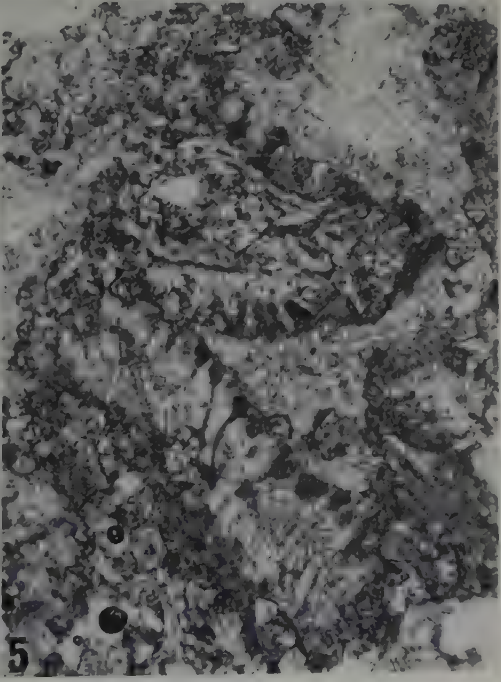
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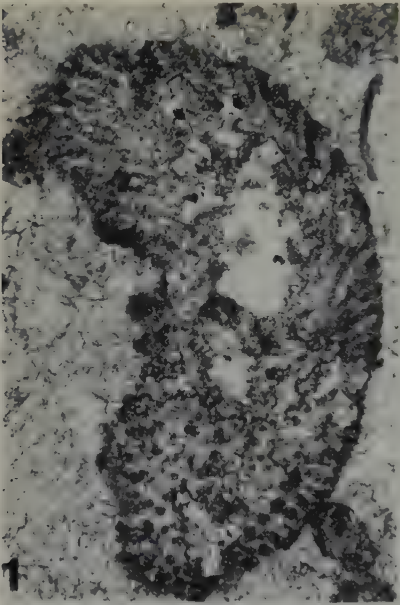

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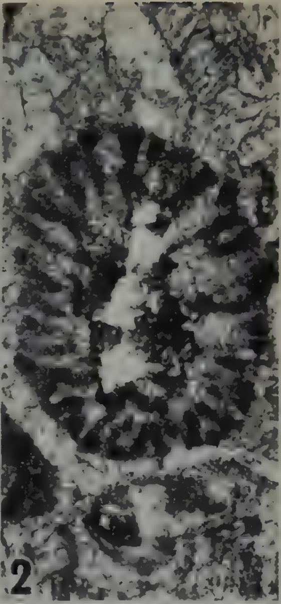

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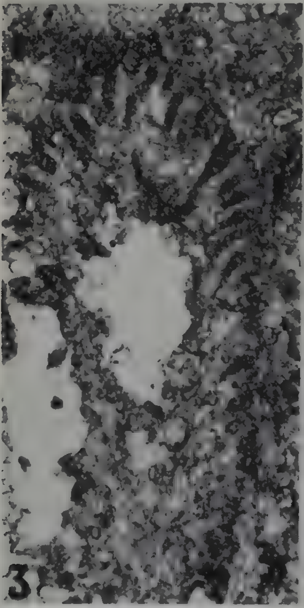

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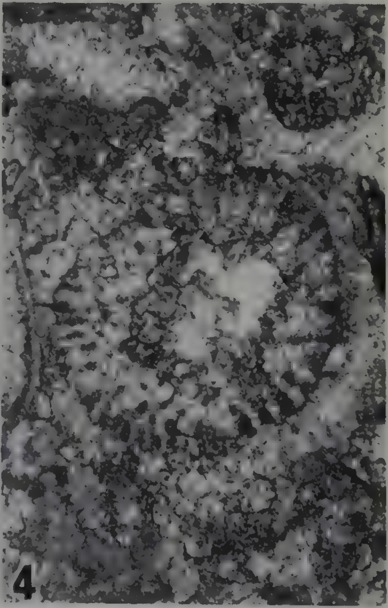

4

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forms of which the Palaeozoic *Ortonella* and Mesozoic *Cayeuxia* are examples, and which may or may not be correctly referred there; and the segmented branching forms of which the modern *Halimeda* is the most familiar. In these latter, which correspond better during life to the popular idea of a sea-weed, and whose segments fall apart after death, there are again two groups. In one, each segment shows coarse longitudinal medullary threads and finer irregularly radial branching subdermal threads: *Halimeda* and *Boueina* (Cretaceous–Recent and Mesozoic respectively; see Pia, 1926) belong here, and this pattern was earlier represented by the Ordovician *Dimorphosiphon* (Hoeg, 1927). In the other, the medullary threads are finer and less distinctly preserved, since the original calcification was patchy and less well developed, and the radial threads, also finer, are longer and of indistinct origin, their terminal portions giving a much thinner and less well-calcified subdermal zone. *Arabicodium* in the Mesozoic (Elliott, 1957) and *Palaeoporella* in the Palaeozoic, are referable here. *Anchicodium* (Johnson, 1946) is an encrusting codiacid from the Carboniferous and Permian, and need not be compared with the segmented Devonshire fossil, and *Orthriosiphon* (Johnson & Konishi, 1956), also Carboniferous, has stumpy club-shaped segments with divergent branching threads set nearly horizontally. Moreover, structures interpreted as internal conceptacles have been described in it: in Recent *Halimeda* the reproductive structures are temporary external outgrowths, of which only one fossil occurrence is recorded (Pfender, 1940), and one would expect *Palaeoporella* to have reproduced similarly. *P. lummatonensis* does not show sporangial cavities or outgrowths on the material examined.

From the type-species *P. variabilis* Stolley (1893), *P. lummatonensis* is distinguished by considerably coarser and more closely set internal branch-detail. Both species are different in segment-shape from the somewhat doubtful *P. grandis* from the Silurian (Stolley, 1896). The Mesozoic *Arabicodium* has much smaller segments and finer detail than *Palaeoporella.* The Russian Devonian genera *Abacella*, *Lancicula*, *Litanaia* and *Uva*, referred to the Codiaceae, appear quite different to *Palaeoporella* from the descriptions (Maslov, 1956).

Pia (1924, 179) quoted a record of Steinmann of an alga related to *Palaeoporella* occurring in the Middle Devonian of the Eifel; no description or figure was given. In view of the known examples of close correspondence of Devonian facies and fossils between Devonshire and Germany, it seems not unlikely that this may be *P. lummatonensis*. The Devonshire fossil bears no resemblance to the European Middle Devonian dasyclad *Zeapora* (Flügel, 1959).

The only other associated alga at Lummaton was a tiny nodule of *Girvanella*-tubules.

All but two of nearly forty sections of *Palaeoporella* occurred in one hand-specimen of limestone collected by my son, Michael Elliott, when assisting me to collect brachiopods, and his help is gratefully acknowledged.

3. PALAEOECOLOGY OF THE LUMMATON SHELL-BED

The Lummaton Shell-Bed is a small portion only of the series of Devonian limestones exposed in the disused quarries at Lummaton Hill, and is recorded in two only of these quarries. It is especially rich in small brachiopods, and many of these were collected by the Rev. Whidborne in the nineteenth century and described by Davidson (1864–82) and by Whidborne himself with the rest of the fauna (1889–1907); he found one of these patches or exposures to be relatively rich in trilobite remains. Whidborne considered the bed to represent a probable current-accumulation and also described it as a littoral deposit (1895, 180; 1889, 2). This second description was criticised by Jukes-Browne (1906, 291–2), who however agreed with the first opinion. The relation of this bed to the rest of the sections is discussed below: an attempt is first made to evaluate Whidborne's suggestion as to its origin.

All of the material for this investigation was taken from the exposure at the top of the eastern face of the small western quarry (Jukes-Browne, 1906, 294). This bed is there confined to this spot and appears to be a stratigraphic entity, and not a portion of limestone individualised out by surrounding dolomitisation. It does not show a high dip, at any rate in the plane intersected by the quarry face, and portions of fossiliferous limestone at the foot of the exposure lay loose and had fallen from above, as indicated by the distinctive weathering. In thin-section it shows a distinctive bryozoan microfacies, described below (p. 255).

From as little as a foot below this level, thin-sections of the limestone reveal a different microfacies, in which small bryozoan- and shell-fragments are uncommon, and missing from some samples. Large pieces of crinoid and stromatoporoid are set in coarse sparry calcite, and dolomitisation is more advanced than in the shell-bed proper, showing as patches, crusts and replacements which make up an appreciable part of the rock. This is the edge of the main massive limestone below, in which dolomitisation has occurred in patches many feet across, in what is essentially a stromatoporoid-crinoid limestone in the unaltered patches.

From the fossil-bed a sample of between three and four hundred small macrofossils was found to consist largely of brachiopods: these amounted to 94% of three hundred determinable fossils. 87% of the three hundred were rounded or globular, and largely smooth-surfaced or only finely ornamented: 13% terebratuloids, 33% spiriferoids, 37% rhynchonelloids, 4% other fossils (*Conocardium* sp., etc.). Most of these 87% fall in the size-group 10–15 mm. geometrical diameter (transverse measurement irrespective of morphology). The larger fossils were mostly single valves of brachiopods or lamellibranchs, with one or two spreads or fans of bryozoa. A few minute immature brachiopods also occurred.

A series of samples of the matrix of this fossil-bed showed that the rock is in thin-section an organic limestone with bryozoan debris predominating. Rather surprisingly, brachiopod-debris is a subordinate element. The bryozoa-debris, amongst which the lattices of fenestellids are especially conspicuous, is set in a matrix of coarse clear sparry calcite, the crystals showing well-marked cleavage. Small clusters and strings of dolomite crystals occur not uncommonly, usually in association with organic remains. Fragments of brachiopods, mollusca, stromatoporoids, corals and crinoids are recognisable and occasionally those of trilobites and algae, with a few ostracods, but all subordinate to the bryozoa.

This assemblage is interpreted as a current-moved accumulation of available loose objects, all organic, offering similar resistance to the water but of different specific gravities, and hence of different sizes and shapes. The brachiopods, some adults of small species, some young of larger species, were closed globular or rounded fluid-filled objects of about the same size and specific gravity which the current would easily move along the sea-floor, once detached from their place of growth. Those with strong ornament, high areas, etc., or different sizes, would offer a different resistance and be separated. The larger but lighter single valves and trilobite-carapace fragments would be moved by the current according to their shape, and pieces of bryozoan-lattice would also be sorted in this way. The study of Menard & Boucot on the experimental movement of brachiopod shells by water (1951) has been applied by the writer (Elliott, 1956) to various post-Palaeozoic accumulations of this nature, and is relevant here. The explanation given above seems the most likely, but depending upon the combination of grain-size of the bottom material and the strength of the current, it is possible for spherical shells to need a stronger current to move them than is necessary for the movement of associated irregular shells, or the current may even sweep away the inorganic grains and bury or concentrate the shells, which are not moved with the bottom material (Menard & Boucot, *op. cit.*). Whilst it is probably impossible to deduce the exact conditions under which the now lithified Lummaton bed formed, it is clearly recognisable as an accumulation due to current-action.

The evidence therefore confirms the early view of Whidborne, with which Lloyd (1933) agreed, as to the drifted nature of the deposit. The brachiopods and other organisms, attached during life to stromatoporoid bioherms and to coral and crinoid growths, were sorted and accumulated in clear shallow and probably warm waters, judging by the algae, on to a bottom of nearly pure calcareous mud and small organic fragments, to form a local accumulation. Lloyd (1933, 73) suggested that this was a consequence of uplift in late Givetian times.

4. GEOLOGICAL AGE OF THE LUMMATON SHELL-BED

It is well known that although the Devonian system was proposed in the early nineteenth century for certain English strata, and that numbers of its fossils were first described from Devonshire localities by early workers such as Phillips and J. de C. Sowerby, most subsequent classification was based on the German, Belgian and American successions, which are much less complicated tectonically, and whose fossils are better preserved. Consequently in the later nineteenth century classification of the type Devonian was pursued largely by the recognition there of various facies and rock-types known elsewhere, notably by W. A. E. Ussher, the closest resemblance being to the German Rhineland succession. This historical development, ably reviewed by Simpson (1951), affects the present consideration of the age of the Lummaton Shell-Bed.

Ussher (1890, 503) quoting MS. notes of A. Champernowne, referred to the occurrence together in the Lummaton limestones (of which the shell-bed forms but a small part) of the brachiopods *Stringocephalus* and *Hypothyridina cuboides*. In Germany and Belgium these are index-fossils of the Givetian (higher Middle Devonian) and Frasnian (lower Upper Devonian) respectively. He considered that the overlap of their ranges occurred in continuous Givetian–Frasnian sedimentation now represented by the massive limestone (1890, 506; 1903; 1913), and discounted the view of Hughes (Ussher, 1890, 517) that they occurred there at different levels. In Germany the Massenkalk, the equivalent of the Torquay area Massive Limestone, is known to be largely Givetian but to extend up into the Frasnian, and a boundary horizon has been distinguished in one area (Tilmann, 1938, 9, 10, 16, 17).

Whidborne and others were concerned with Lummaton mainly as an unusually rich fossil-locality for the Devonian, but Jukes-Browne (1906) described the succession in the various quarries in considerable detail. In the massive jointed and faulted limestones no definite bedding is discernible, and large portions are completely dolomitised. He traced, however, a general succession through progressively younger beds from south-east to north-west, through the Middle Devonian into a limestone in the westerly part of the northernmost (north-western) quarry yielding the coral *Phillipsastraea hennahi* (Lonsdale), which is characteristic of the limestones at Barton, a detached mass a little to the north of Lummaton. Barton is the type-locality of this species (Smith, 1917), still known locally as 'stars' among older residents formerly familiar with the Barton quarries. In Europe, the species, with others, is characteristic of the Frasnian (Upper Devonian), and on this evidence it seems reasonable to conclude that the Frasnian is represented at Barton and at the extreme north-west at Lummaton. Now if Whidborne's two occurrences of the Lummaton shell-bed are connected on Jukes-Browne's map (1906, 294), the line strikes south-west–north-east, just south of this north-western corner, i.e. at the presumed top of the Middle Devonian. In the eastern occurrence of the shell-bed the fossils were said by Whidborne to occur at top and bottom of the quarry-face, from which he argued a high dip (Davidson, 1882, 6): the double occurrence was confirmed by Jukes-Browne (1906, 292). This is not the writer's conclusion in the western quarry (see above, p. 254) but we have no accurate knowledge of the detailed structure of the hill.

Amongst the brachiopods collected for the present study from the shell-bed were several examples of *Hypothyridina cuboides* (J. de C. Sow.). This species is quoted as a marker for the Frasnian in Germany and Belgium, and at Lummaton occurs in the Massive Limestone with the Givetian marker *Stringocephalus*. The type of this rhynchonelloid species, however, comes from Plymouth (Sowerby, 1840). Now the Plymouth limestones have been zoned on the evidence of corals by Taylor (1951): he concluded that the great bulk of the thickness was Middle Devonian, with a little Frasnian at the top. A comparison of Taylor's stratigraphic table (1951, 53) with Ussher's summary of T. Davidson's and R. N. Worth's statements on Plymouth fossil-localities (1907, 53) suggests strongly that Sowerby's type came from one of the old quarries in the Middle Devonian, though absolute proof is probably impossible, as with so many old collections.

In Germany another species, *Hypothyridina procuboides* (Kayser), was described from the Middle Devonian and distinguished on grounds of form and detail from the Upper Devonian '*H. cuboides*' (Kayser, 1871): this author later recognised *H. procuboides* from Lummaton and Hope's Nose, Torquay (Kayser, 1889, 186). Whidborne (1893, 134, 135) protested, quite correctly, that whatever the validity of the separation of these two species, his Lummaton examples were the true *H. cuboides* (J. de C. Sow.).

Of later German brachiopod studies, much the closest analogy to Lummaton is that of Torley (1934) on the fauna of the Upper Givetian Massenkalk from a Ruhr locality: he himself used Lummaton as a comparison in his species-distribution table. Torley recognised (1934, 87) that Kayser's criteria for delimitation of *H. procuboides* and *cuboides* were invalid if enough material to show variation was studied, but apparently following Paeckelmann (1913, 263) and Leidhold (1928, 46), he used a fine marginal bifurcation or splitting of the ribbing as a characteristic of his Middle Devonian forms, which he determined as *H. procuboides*.

The present writer has examined specimens of large typical '*H. cuboides*' from the Frasnian of Belgium. This material does not show this marginal splitting on shell or cast, though there may be marked chevron growth-lines on the ribbing in this area.

In old collections of this species from England, with the general label 'Lummaton' or 'Barton', both types occur, with those showing splitting in a majority. These are not of course localised in detail with reference to the extensive sections, and Middle Devonian occurs at Barton as well as at Lummaton (see House, 1956, 262). Preservation is poorer in this altered limestone material than in the Belgian examples.

Sowerby's figured type (1840, pl. 56, fig. 24), probably selected as a large and distinctive specimen rather than as an average one, as was then the custom, cannot be recognised amongst the relevant material in the Geological Survey Museum, if it is assumed that his figure was accurate, so cannot be examined for this character.

On general grounds it would seem that marginal splitting of the ribbing, as seen on indifferently preserved specimens, could be a consequence of chevron growth-pattern in the marginal shell, and hence likely to occur in all individuals of this species or species-group. Schmidt (1941, 27) has discussed the relationship and difficulties of distinguishing the two species. However, none of the limited stratigraphical evidence contradicts Torley's distinction, though *H. cuboides* (J. de C. Sow.) is the correct name to apply to the Middle Devonian forms, which include *H. procuboides* (Kayser), and which may be abundant at the junction level, and *H. cuboides* auct. from the Upper Devonian needs a new name. It is hoped that a current revision of these rhynchonelloids by Dr. H. Schmidt will clarify this. *H. margarita*, from the Australian Upper Devonian (Veevers, 1959, 100) shows ribbing of the same type as the European Upper Devonian form, i.e. without marginal splitting.

Of importance here is the agreement of the Lummaton shell-bed examples collected by the writer with Torley's Upper Givetian '*H. procuboides*' from Germany. This strongly suggests that the former bed is Givetian, not Frasnian, and its position in the quarries suggests that it is at or near the top of the Middle Devonian. If Whidborne's little immature *Stringocephalus* (1893, 97) came from this bed it would confirm the age as Givetian: it seems probable that it did, but it was not found in the present writer's sampling. The species is not a common one at Lummaton. *Palaeoporella* (see above, p. 253) is also recorded from a German Middle Devonian level.

A revision of the Lummaton–Barton fauna and the occurrence of its species in the quarries, so far as can be ascertained, is desirable. Meanwhile the evidence points to the age of the shell-bed proper as top Middle Devonian (Upper Givetian).

My thanks are due to the authorities at the British Museum (Natural History) and Geological Survey Museum, for facilities for examining brachiopods in connection with this work.

EXPLANATION OF PLATE 9

All figures of *Palaeoporella lummatonensis* sp. nov., thin-sections from the Lummaton Shell-bed, top Middle Devonian (Upper Givetian); Lummaton Hill (small western quarry), Torquay, Devonshire.

1. Holotype, longitudinal section of incomplete segment, × 12; British Museum (Natural History), Dept. Palaeontology; registered number V.42603.
2. Paratype, oblique-transverse of thick, probably basal, segment, × 12; registered number V.42604.
3. Oblique-transverse of ordinary segment, × 20; registered number V.42605.
4. Transverse cut of small segment, × 28; registered number V.42606.
5. Paratype, random sections to show details of internal structure, × 28; registered number V.42607.

EXPLANATION OF PLATE 10

All figures of *Palaeoporella lummatonensis* sp. nov., locality and horizon as for Plate 9; thin-sections × 28.

1. Paratype, oblique-transverse section of curved segment; registered number V.42608.
2. Paratype, transverse of ordinary segment; registered number V.42609.
3. Oblique-longitudinal of ordinary segment; registered number V.42608.
4. Paratype, transverse of ordinary segment; registered number V.42608.

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Field Meeting on the Lower Tertiaries at Swanscombe, Kent

1 May 1960

Report by the Director: DENNIS CURRY

Received 3 March 1961

THE MAIN PURPOSE of the Field Meeting was to visit the celebrated road-cutting on the Watling Street at Swanscombe Park. This was originally excavated between 1922 and 1924, and the Association visited it twice during that period (Brown & Priest, 1924). After it had been completed, the cutting was grassed over and the sections which it had displayed were covered up. In the early part of 1960, to provide an extra carriageway, the cutting was widened by about a hundred feet on the south side. As a result the magnificent section through the Woolwich and Blackheath Beds, originally described by Brown & Priest, became visible once more.

About thirty members and friends assembled at Swanscombe Halt station and proceeded southwards along the High Street, Stanhope and Southfleet Roads towards Watling Street, passing a large chalk pit on their right-hand side. A short visit was paid to the southern end of this pit in the neighbourhood of map reference 613730.[1] Here was seen an inconstant series of sands and gravels containing Woolwich and Blackheath fossils lying on a very irregular surface of Chalk. The Chalk was exposed to a depth of about thirty feet. Although it contained quantities of flints, these were not disposed in layers but were scattered at random throughout the whole mass of Chalk exposed. The party discussed at length the phenomena revealed and it was decided that the whole thickness of Chalk visible had been much disturbed, while the overlying sands must have travelled downhill some distance from the higher ground around.

A very short distance (at 613729) from this disturbed section was seen a small section in pale, fine sands of the Thanet Beds which showed no signs of disturbance whatever. At the corner of Southfleet Road and Watling Street the Director pointed out a transformer station which had recently been erected and stated that a total thickness of at least thirty feet of typical Thanet sands had been visible during the course of its construction. These also showed no signs of disturbance.

Leaving Southfleet Road, the party proceeded westwards along Watling Street to the Swanscombe Park road cutting (596727 to 603727). Here the following section was seen:

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[1] All grid references lie within the 100 km. square.

			ft.	in.
BASE OF LONDON CLAY	Grey brown clay without fossils	About	2	0
BLACKHEATH BEDS	Yellow and white sands with seams and lenticles rich in battered and well-rounded flint pebbles and molluscan shells	About	12	0
WOOLWICH BEDS	Buff and grey sandy laminated clay		4	0
	Grey clays, becoming sandy in the lowest 3 feet. A discontinuous band of grey-green marl, up to 3 inches thick, about 6 feet from the top. Many fossils		11	0
	Grey sandy clays as above		3	0
	Shelly grey-green sand with *Corbicula*		2	6
	Ferruginous sand with oysters		0	9
	Purple loam, carbonaceous in parts		0	6
	Buff sand	About	5	0
WOOLWICH BOTTOM BED	Pebbly glauconitic sand, seen in floor of road at east end of section	Thickness not ascertained		

The above account is based on a section originally compiled by Mr. S. W. Hester in 1924, with modifications by Mr. Hester and Miss E. W. Andrews in 1960. Thanks are rendered to the Director, Geological Survey and Museum, for permission to publish details from this section.

The fauna of the Woolwich Beds consists almost entirely of molluscs. These are beautifully preserved, but of relatively few species. *Corbicula cuneiformis* occurs in myriads throughout the fossiliferous beds while in certain bands *Corbicula cordata*, *Brotia melanioides*, *Tympanotonus funatus* and *Ostrea tenera* are common. The marl bed in the middle of the main clay seam proved to be rich in very small molluscs, including *Planorbis hemistoma*, *Bithinella websteri* and *Lapparentia pygmaea*. A most interesting record from this bed was the discovery (believed to be the first record from the Woolwich Beds) of fruits of Charophytes. Dr. L. Grambast of Paris kindly reports that these are of a new species of *Stephanochara*, a genus not hitherto known below the Upper Eocene.

The Blackheath Beds have yielded a rich series of molluscs of about fifty species. They are for the most part very fragile and must be transported with care (preferably in dry sand) and be treated with a hardening solution. The commonest species are the lamellibranchs, *Glycymeris plumstediensis* and *Nemocardium plumstedianum*, with *Pitar obliquus* and *Corbicula cuneiformis*. The gastropods *Aporrhais sowerbii*, *Brotia melanioides*, *Sigatica abducta* and *Calyptraea* sp. are not rare. Sharks' teeth also are fairly common.

The Director drew attention to the composition of the fauna of the Blackheath Beds, pointing out that it contains a mixture of marine and estuarine forms and that the proportion of estuarine forms diminishes from west to east. In this respect the fauna at Swanscombe was stated to be intermediate between that of Abbey Wood on the one hand and Upnor and High Halstow on the other.

Members of the party spent some time in examining and collecting from the section in the road-cutting. Afterwards they proceeded southwards through a wood to the face of a quarry at 599725. Here were seen about thirty feet of grey-brown unfossiliferous silty clay forming the lowest beds of the London Clay. It then became apparent that the wood through which the party had walked covered the floor of the quarry, which was last worked by the Associated Portland Cement Company Ltd. over forty years ago. The quarry floor coincided approximately with the base of the London Clay and workings came to an end when further extension would have threatened a reservoir at 599724.

The party walked back to Swanscombe for tea, during which Dr. J. M. Hancock, on behalf of members, thanked the Director for organising the excursion.

NEW FOSSIL RECORDS

A very full list of fossils was given by Wrigley & Davis (*in* Brown & Priest, 146, 147). Most of the species there recorded have been rediscovered. The following have been found in addition.

Woolwich Beds. *Ostrea tenera* J. Sby., *Scrobiculabra condamini* (Morris), *Lapparentia pygmaea* (Morris), *Bithinella websteri* (Morris), *Pollia* cf. *lata* (J. Sby.), *Stephanochara* sp.

Blackheath Beds. *Phacoides uncinatus* (Defr.), *Psammobia* sp. *Mactra levesquei* d'Orb.

ACKNOWLEDGMENTS

Thanks are due to Mr. J. Collins, who first drew attention to the new section, and to the County Surveyor, Kent County Council, and Messrs. T. H. Contractors Hire (Chiswick) Ltd. for permission to visit the site.

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WRITTEN CONTRIBUTIONS TO THE DISCUSSION OF A PAPER PREVIOUSLY TAKEN AS READ

The Editor

Dear Sir,

I would like to comment on a number of points in Dr. Tresise's (1960) interesting paper on the Wessex Upper Greensand.

Tresise suggests well-oxygenated waters are unsuitable for the formation of glauconite. Studies of recent glauconitic sediments indicate, however, that glauconite forms under conditions of good oxygenation, turbulence and slow deposition on continental shelves, swells or banks (van Andel & Postma, 1954; Collet, 1905; Murray & Renard, 1891). However, *local* semi-reducing conditions provided, perhaps, by decomposing organic matter seem to be necessary, because about 12% of the total iron in glauconite is in the ferrous state (Burst, 1958).

Tresise's extension of the term 'metacolloidal' to cover *structures* seems most undesirable; Rogers (1917) advocated using Wherry's (1914) term 'metacolloidal' —signifying microcrystalline *substances* of colloidal origin—to avoid loose usage of the terms 'amorphous' and 'colloidal'.

Finally, I am unable to agree that the bulk of the glauconite is of colloidal origin. Much of the available evidence suggests an origin by alteration of other silicates, mainly clay minerals and biotite. Neither the random aggregate structure nor desiccation cracks in glauconite necessarily imply an origin involving a colloidal or gel stage.

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The Editor

Dear Sir,

There are several aspects of Dr. Tresise's (1960) interesting paper with which I must disagree.

It is incorrect to assume that the base of the Chalk can be taken as approximating to a time-plane (p. 334), for over most of Dorset the basement bed is of Middle Cenomanian age, and not Lower Cenomanian as elsewhere, and so is also diachronous. Because of this, the Upper Greensand over parts of Dorset was exposed longer and suffered a greater amount of erosion.

Allowance has not been made for thickness variations of Upper Albian zones, so that lithologies of different ages are shown upon the same map. For example, in the Isle of Purbeck the dispar zone is condensed, six to ten feet of greensand being equivalent to thirty + feet of Upper Greensand in the Vale of Pewsey and SE. Devon. Fluctuations in other Upper Albian sub-zones probably also occurs (Wright, 1947), and some may even be missing (Wilson *et al.*, 1958). Consequently, the maps of Fig. 6 are not lithofacies maps for they do not show lateral lithological variations within time-rock units, but merely show the lithologies across arbitrary diachronous planes.

In map 5, Fig. 6, the pebbly grit of W. Dorset is the top of the 'Eggardon Grit', the occurrence in which of Cenomanian ammonites (Jukes-Browne, 1903; Wilson *et al.*, 1958) suggests it should be correlated with the Al limestone of SE. Devon rather than with the Upper Greensand. If this is so, then the 'belt' of grit becomes the eastern edge of a sheet of grit spreading from Devon into Dorset.

The attempt to represent two independent and variable factors of the Chert Beds on one map (Fig. 8, p. 338) has resulted in over-generalisation, for the accurate presentation of one factor has often led to inaccuracies in the other, giving a false impression of constancy to the two factors for each region.

Finally, on p. 338 Tresise quotes King (1954) as evidence that Brittany and Cornwall were united in the Upper Albian. Although King's terminology is rather confusing, it is clear that he includes the Upper Greensand with the Upper, and not the Lower Cretaceous, and in fig. 2, p. 86, he shows two records of Upper Greensand from the Western Channel south of Plymouth.

It thus seems likely that the Upper Albian sea penetrated between Cornwall and Brittany into the Atlantic, particularly as this route was probably open during the Aptian (viz. the Iberian affinities of the Punfield Limestone).

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The Editor

Dear Sir,

G. R. Tresise (1960, 316–39) has recently discussed the lithology of the Upper Greensand of the West of England. In his paper he has expressed a number of views that may be questioned. I would like to draw attention here to certain inaccuracies or misleading statements concerning the zonal stratigraphy.

He states (1960, 334) that no detailed zonal scheme has yet been produced for the Upper Greensand. This is quite incorrect. Spath (1941, 668) gave a zonal scheme for the Albian which, with little modification, had been used throughout his Monograph of the Ammonoidea of the Gault in which ammonites from the Upper Greensand of the West of England had been described. C. W. Wright has applied this zonal scheme to some of the Dorset coast sections (1947) and it has and can be applied elsewhere. The zonal and sub-zonal classification of the Upper Albian given by Spath in 1941 is as follows:

ZONE	SUB-ZONE
Stoliczkaia dispar	*S. dispar* & *Mortoniceras* (*Durnovarites*) *perinflatum*
	Arrhaphoceras substuderi
Mortoniceras (*Mortoniceras*) *inflatum*	*M. inflatum var. aequatorialis*
	Callihoplites auritus
	Hysteroceras varicosum
	Hysteroceras orbignyi

The *dispar* Zone is represented by the Ammonite Bed *beneath* the Chert Beds on the Dorset coast from Ringstead Bay to Swanage Bay. The Eggardon Grit which Mr. Tresise classifies with the '*Pleurohoplites dispar-perinflatum* Sub-zone' is at least in part contiguous with the Ammonite Bed on the Dorset Coast. The Chert Beds of the Isle of Purbeck coast sections represent a gap between the *dispar* Zone and the *varians* Zone Glauconitic Marl and may be uppermost Albian and basal Cenomanian in age. They are in any case later than the *dispar-perinflatum* Sub-zone ammonite bed, not earlier. It is pertinent to mention that there is evidence to show that the *dispar* ammonite bed is diachronous containing later material eastwards. Apart from the species of *Schloenbachia* recorded by Wright (1947, 184, 186) from Punfield, the writer has obtained a small *Submantelliceras* from the *dispar* bed in Lulworth Cove. At Holworth House, Bed 9 (Wright, 1947, 191) has yielded poorly preserved ammonites which indicate the *substuderi* Sub-zone. Without doubt the Malmstone includes deposits of both the *aequatorialis* and *auritus* Sub-zones. It is unfortunate that Mr. Tresise should have apparently overlooked the importance of the presence of the *dispar-perinflatum* Sub-zone Ammonite bed in his paper. The Upper Greensand was deposited in much shallower water than the contiguous Upper Gault of eastern England. Certain depositional features apparent in the upper part of the Lower Gault and in the Upper Gault of eastern England have a marked effect on the sediments of Upper Albian age in the west of England. Local conditions of deposition also varied considerably, and so similar rock-types might have been formed which are not of the same age, although the vertical sequence may seem to be superficially similar.

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H. G. OWEN,

54 Auckland Rise, London, S.E.19.

The Editor

Dear Sir,

I am very sorry that Dr. Tresise's useful paper, bringing out the essential and distinctive features of the Upper Greensand, was not read before publication. In his first sentence it is stated that the Upper Greensand is '. . . of Upper Albian age', yet even if one excludes the sandy beds of Cenomanian age in Devon, the conventional Upper Greensand of west Dorset includes the Eggardon Grit (Wilson *et al.*, 1958) whose upper part is Cenomanian. Similarly, the captions to the lithofacies maps are misleading because presumably they should read 'n feet below the base of the Cenomanian'. In No. 5, for example, '1 foot below the Chalk' places the horizon at Middle Cenomanian in parts of Devon and Upper Albian in east Dorset.

It has been shown recently (Hancock *in* Mottram *et al.*, 1957) that the base of the Upper Greensand in south Wiltshire belongs to the *varicosum* or *auritus* Sub-zone, and hence is probably the same age as the Blackdown Greensand. Therefore the initial development of Upper Greensand in Wessex extended farther to the east than suggested by the author.

The nodules in the upper part of the formation contain less glauconite than the surrounding sediment although the glauconite has not been corroded by the calcite cement. This suggests that the smaller quantity of quartz in these nodules is not solely due to calcite corrosion but is also an original feature.

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J. M. HANCOCK,

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Studies of glauconite formation in recent marine sediments are not as unanimous in their conclusions as Mr. Zumpe would imply. For example, Takahashi (1939) postulated a strongly reducing environment and Cloud (1955) a weakly reducing one. The evidence of the Upper Greensand suggests that, while strongly reducing conditions are unnecessary, strongly oxidising conditions are definitely unsuitable.

'Metacolloidal' certainly implies a particular mode of origin, and in the laboratory could be confined to substances whose actual process of formation has been observed. In the case of natural materials, however, it must inevitably be applied because of the micro-crystalline structure—the visible end-product from which the mode of origin is deduced.

Positive evidence of the operation of colloidal solutions in glauconite formation is provided by limestone pebbles permeated with glauconite, and by green-stained quartz grains. The mode of origin favoured by Mr. Zumpe does not necessarily exclude the possibility of a colloidal phase. Takahashi (1939) gave a long list of possible source materials, including clay minerals and biotite, but claimed that in all cases glauconisation was accompanied (or preceded) by gelatinisation.

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Mr. Drummond rightly stresses the limitations of the maps shown in Fig. 6, since the Upper Greensand does not represent a chronological unit. It can, however, be regarded as a lithological unit, and was treated as such in Fig. 6 as throughout the paper. Similarly the Eggardon Grit was considered as a unit, since the main depositional break occurs above it, and since Cenomanian ammonites occur only in the highest part, the remainder yielding Albian forms.

The purpose of Fig. 8 was to summarise variations shown in more detail in the columnar sections of Figs. 3 and 5; some generalisation was unavoidable.

I must apologise for misinterpreting King's work which, in fact, neither proves nor disproves the presence of Upper Greensand in the Western Channel. He distinguished undoubted Upper Greensand (recorded only off the Dorset coast) from the glauconitic limestones and cherts of indeterminate age found in mid-Channel farther west. Whether the latter are properly referred to the Upper Greensand of the English Albian or to the Craie Glauconieuse of the French Cenomanian has yet to be established. Nevertheless, Mr. Drummond's suggestion that the sea breached the western barrier during Upper Albian times is obviously a strong possibility.

The correlation of lithological divisions and the Upper Albian zones (p. 334), contested by Mr. Owen, was that suggested by Spath in 1926. The point I attempted to make was that Spath's zonal scheme has yet to be utilised to subdivide the Upper Greensand throughout Wessex as a whole, in the way in which C. W. Wright (1947) had applied it to the Dorset coast sections. Many problems of correlation remain unsolved, but a detailed discussion of these would seem more appropriate in a paper dealing with the palaeontological rather than the lithological aspects of the beds.

Dr. Hancock makes the valid point that the age of the Upper Greensand is not absolutely constant. The 'base of the Chalk' was chosen as datum for the maps of Fig. 6 since the widespread depositional break which separates the Upper Greensand from the overlying Glauconitic Marl is usually coincident with the Albian–Cenomanian junction. Difficulties arise in west Dorset where the break occurs within the Cenomanian, and near Devizes where there is a transitional junction and no obvious break between the two divisions.

The possibility of the basal Upper Greensand in the Vale of Wardour being of *varicosum* age is an interesting one, particularly in view of the apparently diachronous nature of the Malmstone (Fig. 7, 337).

It is difficult to generalise on the nature of the limestone nodules since these show every gradation from cemented aggregates of the surrounding sediment to very pure types virtually free from detrital grains. Thus while nodules rich in glauconite but with few included quartz grains suggest a replacive origin, the purest limestone was probably formed by the recrystallisation of patches in the original sediment rich in organic remains.

ERRATA

Tresise, G. R. 1960. Aspects of the Lithology of the Wessex Upper Greensand. *Proc. Geol. Ass., Lond.*, **71**, 316–39.

p. 334. The table at the foot of the page should read:

Pleurohoplitan	*dispar-perinflatum* sub-zone—Eggardon Grit of western Dorset
	substuderi sub-zone—Chert Beds of western England
Pervinquierian	*aequatorialis* sub-zone—Malmstone of western England
	auritus sub-zone
	varicosum sub-zone—Blackdown Upper Greensand

p. 335, line 5. Delete *Pervinquieria*
line 11. Delete *Pervinquieria*

Strand-line Pebbles in the Mid-Hastings Beds and the Geology of the London Uplands.

Carboniferous Pebbles*

by P. ALLEN

Received 20 July 1960

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ABSTRACT: Abundant Lower Carboniferous cherts from the London Uplands are found in the Top Lower Tunbridge Wells Pebble Bed and the Top Ashdown Pebble Bed. Those dated precisely prove to be Viséan. Nothing older than Middle Viséan is known. The great majority are silicified limestones, representing many of the micro-lithologies found in the massif (=block) facies of the English Viséan. Pebbles of silicified reef limestone may also be present.

Of interest to Lower Carboniferous palaeogeography is the suggestion that at least part of the London Platform was submerged during Upper Avonian times. So far, Carboniferous rocks have not been found in the Platform.

Some of the Carboniferous cherts are impregnated marginally with glauconite, possibly having been derived via Jurassic lydite beds.

Upper Carboniferous materials, if present, are very scarce.

1. INTRODUCTION

ALL THE *undoubted* Carboniferous materials in the Hastings pebble beds are sedimentary cherts ('lydites') of Dinantian age. This was suspected long ago by Milner (1923, 298). Where greater precision is possible they prove to be Viséan. Nothing earlier than Middle Viséan (upper C_2S_1–S_2) or later than D_2 is known. There is no recognisable trace of Tournaisian. Upper Carboniferous may be represented by rolled pieces of silicified wood, but these could be older or younger than that.

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* Jurassic pebbles have been reported previously (Allen, 1960). Old Red Sandstone and other components will be described later.

2. LOWER CARBONIFEROUS PEBBLES

(a) General Petrography

Probably all the tough and compact cherts (Allen, 1960, 163) described below are Avonian. Usually they provide unambiguous evidence of formation from the silicification of pre-existing limestones. In most cases silicification is complete, traces of remanent carbonate being rare. Only one pebble of actual limestone has ever been recorded (Milner, 1923, 288, 297 and pl. 21A). Terrigenous impurities (detrital quartz and clay) have resisted chertification. Some of the cherts are impregnated marginally with fresh glauconite (12 out of 81 chert slices from the Top Lower Tunbridge Wells Pebble Bed and 3 out of 137 from the Top Ashdown Pebble Bed) or collophane. This, with their noticeably greater rounding and less frequent patination than the Portlandian cherts, could suggest secondary derivation through a marine Jurassic intermediary—presumably the Lydite Beds. However, interstitial glauconite is known in the cements of both horizons. Registered numbers (Reading University Geology Department catalogue) of pebbles showing glauconite impregnation are: S1031B1, S1040B1, S1043B1, S1057B1, S1081B1, S1085B1, S1099B1, S1101B2, S1102B2, S1104B1, S1121B1 and S1126B1 from the Top Lower Tunbridge Wells Pebble Bed, and S5005B1, S5089B1 and S5139B1 from the Top Ashdown Pebble Bed.

A bewildering array of former limestones is represented. Most of the main types known in the Avonian of southern England and the Pennines seem to be recognisable. Thin sections viewed in ordinary light often show the original sedimentary structures very clearly. So perfect can the preservation be that the rock is sometimes indistinguishable from unaltered limestone without further examination (Plate 11). Fossils, original lamination, clastic and bioclastic textures, faecal pellets, algal nodules, ooliths, calcite cleavage (S1037B1), carbonate rhombohedra, cavity fillings, etc., are often faithfully traced by the distribution of impurities and/or the varying fabric and composition of the replacing silica. Frequently, too, the sedimentary structures stand out in relief on the surfaces of the pebbles, owing to differential attrition during transport. But in many, of course, the origina features have been entirely obliterated. Signs of deformation or shattering of the limestones, such as stylolites (S1065B1 (red-stained)) and brecciation (S1013B1, S1076B1, S5059B1), are unusual.

The secondary silica is normally quartz or length-fast fibrous chalcedony Other forms (quartzine, lutécite) occur (e.g. S5172B1) but are rare. No further mineralogical details will be given here. In colour the pebbles vary from white through grey to black or brown, and may be mottled with any of these shades. When the white colour is a true patina it often penetrates to the centre of the pebble. Its origin could be partly very recent, for the proportion of patinated cherts (all types) seems to rise in some of the older quarries (cf. Hook Quarry (West Hoathly) with Paxhill).

(b) Top Lower Tunbridge Wells Pebble Bed

The types of chert described below are scattered along the eroded north–south distributary tract (Allen, 1959, fig. 12). All have been found at both West Hoathly (Hook and Philpots quarries) and Paxhill; but, unless otherwise stated, only typical examples from Paxhill are quoted by catalogue number.

(*i*) *Silicified Algal Limestones*. These include former 'concretionary' limestones comprising algal nodules up to one or two inches diameter set in a more argillaceous matrix (S1074B1), former limestones with traces of algal tubes but no true nodular structure (S1069B1, S2011B1 (Hook Quarry)), and former 'mixed' limestones consisting of scattered clusters of algal tubes with other fossils such as foraminifera, crinoidal debris, etc. (S1121B1). Quartz veining is known (S1069B1). According to Professor Alan Wood some of the nodules are best described as spongiostromid, being 'the typical Girvanella of many American authors, layered, without any internal structure' (S1074B1). 'Undoubted *Girvanella*' (S1071B1) and tubes to which the 'closest match is *Ortonella*, from the Carboniferous' (S1069B1) were also recognised by him. Regarding the latter, Professor Wood goes on to say 'unfortunately the French have described a similar form from the Jurassic and the preservation is not good enough to decide which it is'.

A few sections show traces of possible dasycladaceous algae (S1118B1). Dr. R. H. Cummings suggests that certain structures in S1036B1 and S1043B1 'may be highly altered, poorly preserved coccoliths'.

On the fossil evidence alone these cherts might well belong to any system from Silurian to Jurassic. Their petrographical characters, however, link them closely with types containing undoubted Lower Carboniferous fossils (see below).

(*ii*) *Silicified Foraminiferal Limestones*. Pebbles of foraminiferal chert must be counted among the delights of Wealden petrology. Some, particularly a brown-and-grey or grey-and-white mottled form, are recognisable in the field. Their surfaces reveal the larger foraminifera, up to 1 mm. across, preserved in dark silica and clearly visible to the naked eye (S1101A1–S1105A1 (Plate 11, Figs. F, I, J), S3007A1 (Philpots Quarry)).

The following fauna, obtained from Paxhill, was identified by Dr. R. H. Cummings:

Tetrataxis conica: S1103A1 (Plate 11, Fig. F)
Tetrataxis cf. *conica:* S1103B1, S1103B2
Tetrataxis cf. *decurrens* (Brady): S1101A1
Tetrataxis spp.: S1041B1, S1102A1 (Plate 11, Fig. I), S1103B1, S1103B2, S1104A1
Plectogyra cf. *bradyi:* S1091B1

Plectogyra spp.: S1091B1, S1105A1 (Plate 11, Fig. J), S1105B1, S1121B1
Bradyina sp.: S1091B1
Earlandinella spp.: S1091B1, S1105B1
Endothyranopsis sp.: S1103B1, S1103B2
Archaediscus karreri sensu stricto Brady ('not the *A. karreri* of many recent papers on Carboniferous stratigraphy' (RHC): S1121B1
Archaediscus cf. *georgei:* S1121B1, S1105B1
Fourstonella fusiformis (Brady): S1121B1
Cribrostomum sp.: S1105B1, S1121B1
Tuberitina sp.: S1105B1
Climacammina sp.: S1041B1
(Also *Hyalostelia* and other sponge spicules, and fragments of ostracods, bryozoa, crinoids and ? brachiopod shells.)

Dr. Cummings further reports that the assemblages in all the pebbles quoted above are Viséan, that of S1121B1 being 'referred with certainty to the D_2 interval of Vaughan', and S1041B1, S1103A1, S1103B1, S1103B2, S1105A1, S1105B1 indicating an S_2–D_2 age.

(*iii*) *Spicular Cherts, Silicified Spicular Limestones.* Monaxon sponge spicules (including *Hyalostelia* (S1101B1–S1101B3) and ? *Reineria*), with diameters ranging up to 140 μ (S1056B1), are thinly dispersed through many of the silicified limestones described here. Their Carboniferous age is often known from other evidence, and when this is so there is no doubt about the original nature of the rock. But occasionally precisely similar spicules make up the bulk of the pebble. This then becomes a 'spongolite', devoid of all other fossils and with no sign of any pre-existing limestone fabric. While unambiguous identification of monaxon spicules is admittedly beset with difficulties (see discussion on p. 280), it seems most likely that (1) these pebbles also were derived from Carboniferous strata, and (2) they were deposited *as chert* rather than as replacements of limestone. Some are quartz-veined (S1043B1).

(*iv*) *Silicified Coral Limestones.* Though corals, when present, show clearly on the surfaces of the pebbles, they have rarely been found. Only two specimens (S1131A1a, S1073A1 (Plate 11, Fig. L)) were suitable for submission to experts. Dr. H. Dighton Thomas reports that both are young rugose corals and generically unidentifiable. Beyond concluding that the rocks are Palaeozoic, therefore, 'it is not possible to say which System'.

(*v*) *Silicified Bryozoan Limestones.* Many of the cherts enclose scattered fragments of bryozoa, which reach 3.5 mm. (S1096B1) to 8 mm. (S4012B1 (Founthill, Newick)) in length in some cases. The fragments are frequently visible to the naked eye in the field (S1119A1), and particularly abundant in pebbles also containing foraminifera (S1103B1, S1103B2, S1104B1). Monaxon sponge spicules are often associated. One characteristic type (S1034B1, S1035B1, S1054B1) is turbid with finely divided (? clay) particles which give the rock a brownish hue in transmitted light.

Identifiable fossils comprise:

Fenestella (?) Stenoporoid polyzoan colony	S1132A1, S1132B, S1132B2 (identified by Dr. K. P. Oakley).

Cf. *Batostomella* s.l.: S1118B1 (identified by Prof. F. Hodson).
? *Clausa:* S1032B1.

Carboniferous source-rocks are indicated.

(*vi*) *Silicified Limestones with Cephalopods.* Traces of nautiloid septate structure occur in several pebbles. One (S1013B1), submitted to Dr. J. Selwyn Turner, 'is certainly an orthoconic nautiloid' and suggested by him as *Michelinoceras* sp. Dr. Turner further adds: 'I feel reasonably confident that the specimen is Carboniferous on palaeontological grounds . . . and . . . although some of the organic material is shattered, it seems quite undistorted—another point in favour of a Carboniferous age.'

(*vii*) *Silicified Limestones with Debris of Ostracods, Brachiopods, Echinoderms.* Cherts of Carboniferous aspect with abundant echinoderm (S1121B1) or brachiopod debris are surprisingly uncommon. Usually the fragments are small and associated with bryozoa and/or foraminifera (S1103B1, S1103B2, S1121B1, S1122B1, S1126B1). Recognisable crinoidal debris has been seen at Paxhill (S1017B1, S1053B1, S1103B1, S1103B2, S1121B1) and West Hoathly (S2010B1), but this could be Jurassic. Possible brachiopod spines were obtained from Paxhill (S1123B1) and West Hoathly (S2010B1), and shell fragments from Paxhill (S1062B1, S1103B1, S1103B2), West Hoathly (S2010B1) and Founthill, Newick (S4012B1). Ostracods are common (S1077B1, S1101B1, S1101B2, S1101B3, S1102B1, S1102B2, S2010B1).

(*viii*) *Silicified Limestone-Sandstones, Pellet Limestones.* Besides the silicified bioclastic rocks described above, a series of relatively unfossiliferous bedded calcarenites (=in part Cayeux's 'calcaire pseudoolithique' (1935)) is also present among the cherts. These grade into (1) 'clotted' types through increasing indistinctness of the grain boundaries, and (2) oolitic types through the appearance of oolitic 'skins'.

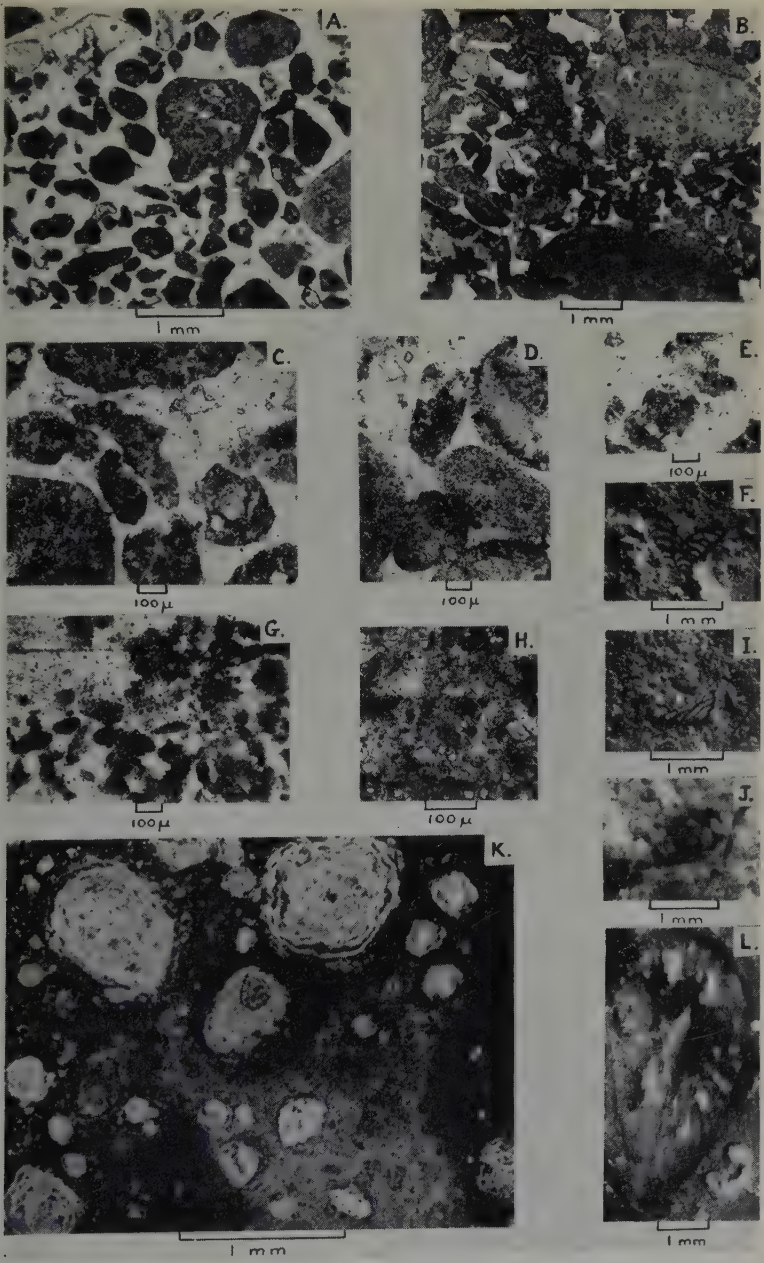
Before silicification most of the rocks comprised rounded grains or pellets of impure fine-grained limestone, usually of sand grade (though occasionally reaching 3–4 mm.) and with sharply defined boundaries. These were set, sometimes 'floating', in a clear carbonate (probably calcite) cement (S1084B1, S4005B1 (Plate 11, Figs. B, A)). Large rounded aggregates of 'clotted' pellets occur at Founthill (S4010B1) and Paxhill (S1098B1). Varieties with pellets of coarse-grained limestone (now quartz-mosaic) are less common (S1085B1). Every sign of rolling and attrition is present. Loose clastic or bioclastic debris is normally absent, only single specimens containing sand grade quartz grains (S1099B1) or macroscopic shell-fragments (S4010B1). However, microfossils such as sponge spicules occur enclosed in some of the fine-grained pellets (S1024B1, S1055B1, S1062B1, S1081B1 (Plate 11, Fig. C), S4005B1) and, less commonly, clear rhombohedra (S1081B1: Plate 11, Figs. C, E). The fine-grained pellets correspond with Cayeux's 'pseudoolithes' (1935, fig. 60a) and 'fausses oolithes' (figs. 59c, 61b, 63b, 64d), being presumably either limestone detritus or aggregates formed during deposition like Illing's composite grains and 'lumps' (1954) and Beales's 'bahamite' particles (1958). The latter are almost certainly represented by occasional large irregular aggregates with embayed outlines (S1081B1 (Plate 11, Fig. C), S1098B1, S4010B1 (Founthill, Newick)). Whether the coarser-grained pellets (='pseudoolithes' of Cayeux's fig. 65) were formed in either of these ways, or bioclastically, or by recrystallisation of former ooliths (cf. Cayeux's fig. 64b, c) is not clear.

Grains other than 'limestone' have not been found. The rocks are obviously more closely allied to the 'clotted' calcarenites and oolites (see below) than to the silicified sandy limestones described later (p. 278). Thus pellets of transitional character may be seen, many with less distinct boundaries (e.g. S1081B1 (Plate 11, Figs. C, E)) like those described by Cayeux (figs. 60a, 63b, 69), and others enveloped in thin oolitic skins (S1084B1 (Plate 11, Fig. D); cf. Cayeux's fig. 59a, c). Mixed varieties comprising jumbled limestone grains (often with oolitic skins) and ooliths are not infrequent (S1040B1). Where the enclosed structures such as microfossils or rhombohedra lie at grain margins they are cut off by the rounded surfaces. Second generations of rhombohedra develop in some of the cements (S1084B1: Plate 11, Fig. D), proving two phases of crystal (dolomite?) growth. The first was presumably intra-Avonian.

Textures identical with the above are known in the Lower Carboniferous limestones of the Pennines and the west of England.

A few (probably $<1\%$) of the pebbles are stained throughout with red ferric oxides (S1062B1). This stain was deposited after the matrix; but whether before or after silicification is not clear. Quartz veins occur in some pebbles, but are uncommon.

(*ix*) *Silicified 'Clotted' Limestones*. As already noted these grade imperceptibly (for example, S1099B1) from the detrital-looking types with rounded and sharply defined grain boundaries. They represent the 'clotted' limestones of Cayeux (1935: 'calcaire à structure grumeleuse') and the 'merged' calcarenites of Beales (1958). Viewed in ordinary light the cherts show every stage from types exactly like Cayeux's intermediate forms (see above), to those with very indistinct grain boundaries (S1003B1 (Plate 11, Fig. H), S1071B1, S1071B2, S2037B1 (West Hoathly)) barely distinguishable from his figures 67 and 68 or Gürich's 'spongiostromid' pellet limestones. Pseudomorphs after rhombohedral carbonate often occur (Plate 11, Fig. H). Closely comparable rocks are known in the C-reef complexes of the British Isles, for example western Eire (Dr. A. Lees, personal communication).


A.
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K.
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[*To face p.* 276

Originally, the roughly spherical pellets (diameters usually <0.2 mm.) seem to have consisted of impure fine-grained carbonate (now fine chert), turbid with dusty inclusions and merging irregularly into the coarser clearer matrix (now coarser silica), the crystals of which often penetrate them. Ill-distinguished pellets thus 'float' in their matrixes with few contacts, giving the typical 'clotted' texture (S1003B1: Plate 11, Fig. H (cf. Cayeux, fig. 67)). Types with irregularly shaped pellets, like Cayeux's figure 68, also occur (S1002B1, S1003B1 (part), S1004B1).

Comparable limestones are known in the C-reef complexes of the British Isles (Dr. A. Lees, personal communication). Probably the original matrixes were not true cements, but fine-grained carbonate (such as calcite-mudstone) which recrystallised prior to silicification.

(*x*) *Silicified Oolitic Limestones.* Beautifully preserved, these grade from the limestone–sandstones or pellet limestones by the development of oolitic 'skins' round the particles of fine-grained limestone. As already noted (p. 276) these particles, the 'pseudoolithes' and 'fausses oolithes' of Cayeux, may have been either erosional in origin or formed by the aggregation of fine carbonate mud during deposition. Other nuclei, such as quartz grains (S1092B1), are very rare. In coarseness, the rocks are always of sand grade, well sorted and virtually devoid of bioclastic or terrigenous debris. Diagnostic fossils have not been found. Current bedding is occasionally recognisable and large rounded grains of oolitic 'limestone' (S1084B1 (Plate 11, Fig. B)) are sometimes seen. The cement, now clear quartz mosaic, seems originally to have been coarse pure carbonate, probably calcite.

While primary structures such as concentric growth attrition zones in the ooliths are normally 'ghosted' exquisitely by the silica, secondary limestone fabrics may also be preserved. Thus under crossed nicols radial structure in the ooliths is frequently detectable, cutting across the concentric zonation seen with ordinary light (S1040B1, S1092B1). This suggests that silicification post-dated recrystallisation of the primary carbonate (aragonite?). Such ooliths are common in Palaeozoic limestones. As already seen (p. 276), some of the coarse-grained pellets in the limestone–sandstones could represent irregularly recrystallised ooliths. With progressive obliteration of the oolitic structure, distinction from the 'clotted' calcarenites becomes more and more difficult. Among all types rhombohedra (suggesting the growth of secondary dolomite) occur, but sparingly (S1084B1).

As with the silicified limestone–sandstones, a few ($<3\%$) of the oolitic cherts are stained red throughout (S1133A1). Again the staining cannot be dated. Milner's record of unsilicified oolitic limestone at West Hoathly has been noted (Allen, 1960, 163), but this was ascribed to a Jurassic source.

(*xi*) *Silicified Fine-Grained Limestones.* Many of the unfossiliferous cherts are so fine grained that thin slices are substantially isotropic at ordinary magnification. Bedding is normally absent, and only irregular cavities and thin wispy veins of clearer quartz (drusy-lined and/or mosaic-centred) occasionally relieve the petrographical monotony (S1047B1, S1095B1).

In some instances the silica has undoubtedly replaced unbedded unfossiliferous limestones of 'porcellanous' type. This calls to mind the calcite- and dolomite-mudstones of the mid-Dinantian reef complexes. At first sight the veined and vesicular varieties enhance the similarity, for they strongly resemble some reef-limestones, including types with '*Stromatactis*' (Bathurst, 1959). However, the interpretation of such fabrics is fraught with difficulty, and at present one cannot be absolutely sure that the vein and cavity fillings are faithful pseudomorphs after similar carbonate structures.

Obviously further evidence is needed before the presence of silicified reef limestone pebbles can be accepted. This may be forthcoming, for in the Ashdown pebble bed there is a suggestion that some of the cavities were floored with spongiostromid (i.e. 'clotted') pellet-rock (p. 281).

(*xii*) *Silicified Rhombohedral* (*? Dolomitic*) *Limestones*. Every gradation is known from all the previous rocks, thinly scattered with silicified rhombohedra (e.g. S1069B1, S1074B1, S1081B1, S1084B1), to types dominated by them. The pseudomorphs are always preserved as inclusion-free silica (chert or quartz mosaic), clearer and coarser than the matrix (Plate 11, Figs. C, D, E, H). Their size is strictly limited, seldom exceeding 50 μ along the larger diagonal, though occasionally reaching 90 μ (S1101B3) or even 0.5 mm. (S1017B1). Bedding is revealed by variations in the distribution of rhombohedra and impurities. The fossils disappear when the pseudomorphs become abundant—a significant point.

Silicified and unsilicified limestones with tiny clear rhombohedra of calcite, dolomite or ankerite are widely dispersed in the Lower Carboniferous rocks of Britain.

(*xiii*) *Other silicified limestones*. While some of the fine-grained cherts give no clear indications of their origin, their fabrics can usually be matched among others of known derivation. Consequently there is no reason to suppose that they came from a different source. A few (S1029B1, S1042B1, S1087B1) are 'sandy', being scattered with angular chips of quartz (cf. S1099B1, p. 275). Others, bearing an obscure brecciated texture (S1076B1), could represent 'pseudobreccia' resulting from the silicification of patchily recrystallised or incompletely dolomitised limestone. Radiolaria (? S1030B1, ? S1056B1, ? S1076B1, ? S1085B1) are unproven, and must in any case be rare.

(c) Top Ashdown Pebble Bed

The striking feature of this horizon is the occurrence of abundant pebbles of the same Avonian cherts as those found in the Tunbridge Wells pebble bed. All the major types described above have been identified, and their relative frequencies are probably about the same as in Tunbridge Wells facies of equal coarseness (table I, Allen, 1960). The overall proportion of chert, however, is probably higher, even in localities farthest from the London Uplands (*op. cit.*).

Owing to wider sampling, the lateral distribution of Avonian materials is known better than in the Tunbridge Wells bed. Again sponge spicules and fragments of bryozoa and ostracods are very common.

(*i*) *Silicified Algal Limestones.* Algal-nodule cherts (Plate 11, Fig. K) are widely dispersed, from the northern limit of outcrop (S5128B1 (Penshurst)) to the southern (S5089B1 (Sedlescombe), S5039B1 and S5042B1 (Iden)). Generic identification has unfortunately proved impossible. The rocks match Avonian types known in the Top Lower Tunbridge Wells Pebble Bed at Paxhill. Professor A. Wood reports: 'The whole appearance is Palaeozoic (Carboniferous or Silurian) rather than Mesozoic.'

(*ii*) *Silicified Foraminiferal Limestones.* Identical with their Lower Tunbridge Wells successors and as widely scattered as the algal cherts, these have been found at Sedlescombe (S5012B1), Iden (S5078B1), Lamberhurst (S5093B1, S5097B1) and Penshurst (S5115B1). The fauna, identified by Dr. R. H. Cummings and Professor A. Wood, is:

Stacheoides sp.: S5115B1
Archaediscus sp.: S5078B1
Tetrataxis cf. *decurrens:* S5093B1
Tetrataxis sp.: S5093B1, S5115B1
Ammodiscid-tetrataxinid types: S5012B1
? Earlandiid: S5097B1
? Valvulinellids: S5093B1
(Also dubious radiolaria, sponge spicules, and fragments of corals, bryozoa, crinoids, ? molluscs and ? brachiopods.)

Dr. Cummings reports that pebbles S5078B1, S5093B1 and S5115B1 are definitely Viséan in age, probably Middle Viséan (= upper C_2S_1 to S_2 of Vaughan's zonal scheme, or F.1.A.4–F.1.A.6 of Cummings's stratigraphical units).

Pseudomorphs after rhombohedral carbonate occur in most specimens.

(*iii*) *Spicular Cherts, Silicified Spicular Limestones.* Cherts, with monaxon spicules of greatly varying size (up to 185 μ diameter (S5111B1) and 1 mm. long (S5022B1)) mixed with other fossil debris, are very common. They clearly represent silicified limestones and are referred to under other headings. The 'spongolite' pebbles, however, are again sharply distinct. These contain only spicules, and no gradations have been found to connect them with the other types of chert. There is no evidence that they were once limestones. Usually well bedded, their near-isotropic matrixes are suffused with either brownish (phosphatic?) patches (Boarzell Hill, Hurst Green (S5021B1)) or irregular areas of included 'dust' (S5012B1 (Sedlescombe), S5024B1 (Darwell Hole), S5131B1 (Penshurst)). The individual spicules are smaller than most of those in the other petrographic types (diameters 10 μ–50 μ, lengths up to 400 μ). Only one specimen dominated by large spicules (diameters around 90 μ) has been seen (S5148B1 (Wadhurst)). Rhombohedral pseudomorphs normally occur, detrital quartz and iron ores are present in a specimen from Hurst Green (S5021B1), and quartz at Penshurst (S5131B1). Colloform structures can be seen to originate in the chert from the sites of spicules (S5084B1).

Detailed investigation by Drs. S. H. Straw, K. P. Oakley and H. W. Ball, including direct comparison with type materials (British Museum (Nat. Hist.) reg. nos. S4225, S4226) shows that the small spongolite spicules belong either to *Reineria zitteli* (Lower Carboniferous) or *Spongilla purbeckensis* (Lower Purbeck). The Boarzell Hill slices can, indeed, be closely matched both petrographically and palaeontologically with material from the Purbeck (Dorset Chert Beds) as well as with Viséan cherts (western Eire).

A Purbeckian age would raise difficult problems. But it is worth noting that (1) Boarzell Hill and Darwell Hole are on the flanks of the Sussex Purbeck inliers, (2) an unconformity between Wealden and Purbeck towards Heathfield is not impossible, and (3) *S. purbeckensis* has been reported in pebbles from the Lower Greensand of West Malling (Kirkaldy, 1947, 235). Rhombohedral pseudomorphs do not necessarily exclude a Purbeck origin either, for both dolomite and chert are developed in the lower beds at the Mountfield gypsum mine. However, the occurrence of similar-sized monaxons in other pebbles carrying datable marine fossils suggests that the spongolites are Avonian. At present this intriguing problem must be left here.

(*iv*) *Silicified Bryozoan Limestones.* As in the Lower Tunbridge Wells pebbles, fragments of bryozoa are among the commonest fossils. They may be collected at most exposures, particularly Hastings (S5003B1, S5004B1), Brede (S5007B1), Guestling (S5008B1), Iden (S5034B1, S5051B1), Lamberhurst (S5097B1, S5099B1, S5101B1), Penshurst (S5108B1, S5115B1, S5120B1) and Stone (S5170B1). Some of the fragments at Penshurst reach 5 mm. in length (S5120B1).

(*v*) *Silicified Limestones with Debris of ? Corals, Ostracods, Brachiopods, Pelecypods, Cephalopods, Echinoderms.* Though various mixtures of these materials (lengths up to 0.65 mm.) and other fossils occur (Iden (S5034B1, S5080B1), Hastings (S5005B1), Holtye (S5138B1), Cowden (S5144B1), Penshurst (S5119B1) and Lamberhurst (S5099B1)), they have never been dated directly.

(*vi*) *Silicified Limestone–Sandstones, Pellet Limestones.* All the petrographic types known in the higher horizon (p. 278) have been found and may be collected from most exposures. Again fine-grained pellets are the rule, though grains of coarse 'limestone' are known (S5104B1, S5129B1

(Penshurst)). Thin oolitic skins surround some of the pellets at Penshurst (S5104B1, S5117B1), Sedlescombe (S5087B1) and Robertsbridge (S5029B1: cf. Cayeux, 1935, fig. 59a, c), and possible algal overgrowths at Wittersham (S5015B1). Unidentified microfossils occur within a large grain (3.5 mm. × 2.5 mm.) at Penshurst (S5113B1), and sponge spicules *between* them at Hurst Green (S5020B1) where there are also two generations of rhombohedra. One specimen (S5101B1) from Lamberhurst contains bryozoa and rectangular pellets. Red-stained varieties have not been found.

(*vii*) *Silicified 'Clotted' Limestones.* These are identical with those in the Lower Tunbridge Wells pebble bed (p. 276) and normally occur as thin courses in otherwise fine-grained unfossiliferous chert (see below). Slices from Lamberhurst (S5100B1: Plate 11, Fig. G) and Sedlescombe (S5088B1), when viewed in ordinary light, can be closely matched with bedded limestones in the European Waulsortian reefs (e.g. western Eire (Dr. A. Lees, personal communication)). Indeed the clotted bed in S5088B1 strongly resembles Gürich's spongiostromid 'genus' *Chondrostroma* (1906, pl. XII, fig. 3).

(*viii*) *Silicified Oolitic Limestones.* Oolites are universal. Gradations from uncoated 'limestone' nuclei have been noted above (S5087B1, S5029B1, S5104B1, S5117B1). Other nuclei, such as shell chips (S5102B1 (Lamberhurst)), are rarely utilised.

Indications of post-recrystallisation silicification (pseudomorphing radial carbonate) are seen at Sedlescombe (S5087B1) and Lamberhurst (S5094B1, S5102B1). Advanced cases of obliteration of the oolitic structure, with the development of chalcedony, quartzine and lutécite, are known (S5172B1 (Stone)). Red-stained oolitic chert occurs at Holtye.

(*ix*) *Silicified Fine-Grained Limestones.* Many of the fine-grained unfossiliferous cherts seem to have replaced calcite- or dolomite-mudstones. Veins and cavities lined with clear fibrous silica and centred with quartz-mosaic (S5033B1, S5082B1 (Iden)), or having fibrous and mosaic silica arranged in alternate layers (S5126B1 (Penshurst)), recall certain reef-limestones of mid-Avonian age (cf. Bathurst, 1959). The specimens from Penshurst (S5110B1, S5126B1) nearly clinch this identity, for their cavities also contain 'clotted' limestone or 'merged' pellets (cf. Bathurst, 1958, pl. I, fig. 4). Rhombohedral pseudomorphs are common.

(*x*) *Silicified Rhombohedral (? Dolomitic) Limestones.* Striking features are the 'picture-frame' zoning of some of the pseudomorphs, seen as planes of inclusions paralleling the former rhombohedron faces (S5027B1 (Robertsbridge), S5002B1 (Brede), S5005B1 (Hastings)), and good bedding, including crossbedding (S5001B1 (Brede)) and graded bedding (S5032B1 (Iden)). The laminae normally appear as courses of different-sized rhombohedra or as planes of impurities (S5017B1). Sometimes rhombohedra transgress the latter.

Traces of fossils other than spicules were seen at Guestling (S5008B1), Hastings (S5003B1, S5004B1, S5005B1) and Sedlescombe (S5012B1). Gradations to spicular chert occur at Boarzell Hill, Hurst Green (S5019B1, S5020B1). 'Sandy' types carrying detrital quartz are known, and one example from Hastings contains grains of muscovite also (S5005B1). Rarely, the rocks were shattered before silicification (S5059B1 (Iden)).

Rhombohedra in a specimen from Hastings (S5004B1) may be pseudomorphic after a more ferriferous carbonate such as ankerite or siderite.

(*xi*) *Other Silicified Limestones.* A feature of the unfossiliferous cherts at the northern limit of outcrop is the frequent occurrence of angular chips of probable detrital quartz (Penshurst (S5106B1, S5113B1, S5115B1, S5129B1, S5131B1, S5135B1), Holtye (S5140B1), Wadhurst (S5153B1)). The grains are about silt-size ($\sim 40\ \mu$). Other 'sandy' varieties with similar-sized grains have already been mentioned above under appropriate headings. Dubious radiolaria are recorded from several localities, but again none is proven.

3. POSSIBLE UPPER CARBONIFEROUS PEBBLES

Pebbles of silicified gymnospermous wood have been collected from both horizons:

Top Lower Tunbridge Wells Pebble Bed
 Paxhill: S1070B1
 West Hoathly. Hook Quarry: S2008B1–S2008B4, S2051B1–S2051B3.
 Philpots Quarry: S3021B1–S3021B3
Top Ashdown Pebble Bed
 Lamberhurst: S5096B1

Professor T. M. Harris, while emphasising the difficulty of distinguishing in single sections between poorly preserved Upper Carboniferous cordaitean and Jurassic coniferous woods, reports that the Tunbridge Wells specimens S1070B1 and S2008B1–S2008B4 (simply organised with apparently no resin or annual rings, tracheids large, rays small and narrow) are, on balance, a little more likely to be cordaitean. Of S2051B1–S2051B3 (orientated sections) he reports: 'This is clearly . . . secondary wood of a member of the Lepidodendraceae . . . large tracheids, no annual rings, simple wood rays . . . scalariform pitting on all tracheid walls . . . [including] . . . tangential walls.'

4. CONCLUSIONS

Cherty Lower Carboniferous limestones outcropped in the London catchment of the 'northern river' during mid-Hastings times. Only Viséan strata appear to have been present. Nothing older than Middle Viséan has been proved.

Massif (=block) facies limestones of considerable purity predominated. Minor developments of reef may have occurred here and there. A wide variety of bioclastic, clastic, precipitated and subsequently altered limestones was represented, and the assemblage can be matched in the autochthonous massif facies of England today. The limestones developed secondary chert, and only chert survives into the Wealden. Pebbles of Carboniferous Limestone are unknown and rocks of basin and Culm facies were absent.

Algae, foraminifera, sponges, bryozoa and ostracods dominated the fossil assemblages. Corals, brachiopods, molluscs and crinoids were not common. Radiolaria are unproven, and must in any case have been rare.

Between top Ashdown and top Lower Tunbridge Wells times the outcrops of Carboniferous Limestone contracted appreciably. The loss through denudation was not entirely made up by the creation of new exposures resulting from removal of the Upper Jurassic cover. At present Carboniferous rocks are unknown in the Platform.

The above picture of the Dinantian element in the Wealden palaeogeology of the London Uplands seems likely to be correct in essentials, despite the possibility that some of the pebbles came secondarily through Upper Jurassic lydite beds. Similar types of chert pebble are known to be abundant in the Upper Lydite Bed near Oxford on the other side of the London Platform (Neaverson, 1925; author's work in progress), and they too are often impregnated with glauconite and/or collophane. But their main significance may lie in derivation from a common source. Thus chert pebbles are absent from the horizon recognised on the south of the Platform at Warlingham (Allen, 1960, 162); and the upward changes in the Hastings pebble suites would not make sense if they were considered to be *wholly* redistributed Lydite Bed materials.

Apparently (1) the bulk of the Wealden cherts came directly from Avonian outcrops on the London Uplands, and (2) a smaller proportion may have come secondarily via Upper Jurassic beds fringing the same source rocks. Wealden denudation continued the work that had been operative since early Mesozoic time, and what had been the London promontory or island in the Jurassic seas now became an internal oldland. In this connection it is interesting to note that both the Upper and Lower Lydite Beds of Oxfordshire have less Old Red Sandstone debris in them than the lowest Wealden (Ashdown) pebble beds.

At present there is no proof that Upper Carboniferous rocks outcropped on the Uplands. In any case their extent would have been small.

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EXPLANATION OF PLATE 11

Numbers refer to Reading University Geology Department catalogue. 'B' indicates thin section. 'A' indicates untreated surface of hand specimen.

T.L. = transmitted unpolarised light. R.L. = reflected light. TLTWPB = Top Lower Tunbridge Wells Pebble Bed. TAPB = Top Ashdown Pebble Bed.

Silicified Calcarenites (*Bahamites?*)

A. Limestone–sandstone or pellet limestone containing sharply bounded grains of former fine-grained limestone (now chert) 'floating' in clear ? calcite cement (now quartz–mosaic). Some grains enclose sponge spicules and rhombohedral pseudomorphs (clearer silica). T.L. S4005B1. TLTWPB. Founthill, Newick, Sussex.

B. As above, but more closely packed. Large composite grain contains ooliths and fine-grained pellets. T.L. S1084B1. TLTWPB. Paxhill, Sussex.

C, E. Transitional type showing pellets with less distinct and more embayed boundaries than A, B. Some enclose rhombohedral pseudomorphs (C, top left and bottom right; E, bottom middle). Others contain sponge spicules (C, south-east centre). T.L. S1081B1. TLTWPB. Paxhill, Sussex.

D. Transitional type showing pellets with less distinct boundaries than A, B. One grain (top right) has a thin oolitic skin. Second-generation rhombohedral pseudomorph developed in cement (top left-hand corner). T.L. S1084B1. TLTWPB. Paxhill, Sussex.

G. 'Clotted' limestone showing change from transitional texture in south-west quadrant to typical 'merged' texture in north-east. T.L. S5100B1. TAPB. Lamberhurst, Kent.

H. As above, but with 'merged' texture only. Rhombohedral pseudomorphs present. T.L. S1003B1. TLTWPB. Paxhill, Sussex.

Silicified Foraminiferal Limestones

Identifications by Dr. R. H. Cummings.

F. Natural surface of chert pebble with *Tetrataxis conica*. R.L. S1103A1. TLTWPB.

Paxhill, Sussex.

I. As above, with *Tetrataxis* sp. R.L. S1102A1. TLTWPB. Paxhill, Sussex.

J. As above, with *Plectogyra* sp. R.L. S1105A1. TLTWPB. Paxhill, Sussex.

Silicified Algal Limestone

K. With small algal nodules. T.L. S5089B1. TAPB. Sedlescombe, Sussex.

Silicified Coral Limestone

L. Natural surface of chert pebble with young rugose coral. Identified by Dr. H. Dighton Thomas. R.L. S1073A1. TLTWPB. Paxhill, Sussex.

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APPENDIX

NATIONAL GRID REFERENCES OF LOCALITIES CITED IN THE TEXT

Locality	National Grid Reference (100 km. square (51) TQ)
Top Lower Tunbridge Wells Pebble Bed	
Hook Quarry, West Hoathly	355313
Philpots Quarry, West Hoathly	355322
Paxhill	356272
Founthill, Newick	426195
Top Ashdown Pebble Bed	
Penshurst	513429
Sedlescombe	785172
Iden	917249
Hastings	781108
Lamberhurst	656367
Boarzell Hill, Hurst Green	722287
Darwell Hole	696187
Cowden	469408
Brede	831184
Guestling	826130
Robertsbridge	721235
Holtye	469391
Wadhurst	661311
Stone	916301
Wittersham	913267

Some Aspects of the Drift Deposits and Geomorphology of South-East Hertfordshire

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ABSTRACT: During the field work of a soil survey of 3rd Edition Sheet 239 (Hertford) it became evident that the origins and distribution of the drifts of south-east Hertfordshire offered a number of problems that have been incompletely solved. In this account further data on the lithology and mineralogical composition of the deposits are presented and their chronology and derivation are discussed. Attention is drawn to the distribution of asymmetric valleys in this and neighbouring areas, and their origin is discussed.

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1. INTRODUCTION

THIS ACCOUNT of the drifts is based on a Soil Survey of part of O.S. Sheet 239 (Hertford). The mapping was carried out with a three-foot soil auger, but in order to appreciate more fully the relationships of the drifts a number of deeper borings were made. The accompanying map (Fig. 1), derived mainly from Soil Survey data, shows the generalised distribution of superficial deposits in the principal area studied, including parts of the parishes of Hatfield, Essendon, Little Berkhampstead and Northaw. Most of the area is drift-covered, but solid formations outcrop in the valleys. Reading Beds and Upper Chalk appear on the northern edge of the map and in the Cuffley Brook Valley, but London Clay underlies most of the area. The upper clayey strata of the Reading Beds in this district weather to a clay subsoil which is virtually indistinguishable from that derived from London Clay, and the two beds are accordingly grouped together on the map as 'Eocene Clay'. The lower sandy facies of the Reading Beds gives rise to easily recognisable soils and is mapped separately.

The interfluves rise to a little above 400 feet and a general plateau surface, sloping gently towards the east and underlain by a varying thickness of drift, can be distinguished. The division of the drifts into Pebble Gravel and Boulder Clay as shown on the published Geological Survey map (New Series, No. 239) was found unsatisfactory when checked by detailed auger survey, since much of the area mapped as Pebble Gravel and some small areas mapped as Boulder Clay proved to have a subsoil of pebbly clay which, unlike the greater part of the Boulder Clay, is non-calcareous to considerable depth (at least three feet). This deposit, described in detail below, is referred to on the map (Fig. 1) as 'Pebbly Clay Drift', whereas the stony clays containing fragments of chalk within three feet are distinguished as 'Chalky Boulder Clay'.

2. THE DRIFT DEPOSITS

(a) Description of the Deposits

As the mapping of the deposits shown in Fig. 1 depends on their identification from soil and subsoil characteristics, texture, stoniness and colour are emphasised in the following descriptions. Textures are defined as in the American Soil Survey Handbook.

(*i*) *London Clay.* This tends to outcrop on middle and lower slopes and the surface layers are often contaminated with drift from higher levels containing pebbles and small chips of flint. At eighteen inches in depth the clay is normally stoneless and prominently mottled with grey and ochreous brown. Below two feet mottling diminishes and a uniform brown-coloured clay is found to considerable depths. Occasional pale brown septaria and many fine crystals of gypsum occur below four feet. The dark grey unweathered London Clay may be as deep as fifteen feet below the surface on drift-free sites.

(*ii*) *Pebble Gravel.* The term Pebble Gravel as used on the accompanying map refers to coarse-textured soil and subsoil materials, consisting predominantly of pebbles and sand, which contain clay only as isolated lenses or as a coating to the sand-grains attributable to weathering or soil-forming processes. In many cases the coarse-textured Pebble Gravel material appears to emerge from beneath the Pebbly Clay Drift at the margins of the Plateau and extends well down the valley sides, presumably as a result of solifluction. For this reason, no attempt has been made to map the 'true base' of the Pebble Gravel, but it is clearly discontinuous as on the plateau the later deposits often rest directly on London Clay, and on many slopes the Pebble Gravel as defined here was not found. This discontinuity may be a result of its original deposition in restricted channels or its subsequent removal due to erosion by ice or other agencies. On sloping sites the soil to a depth of fifteen inches or so is usually very pebbly

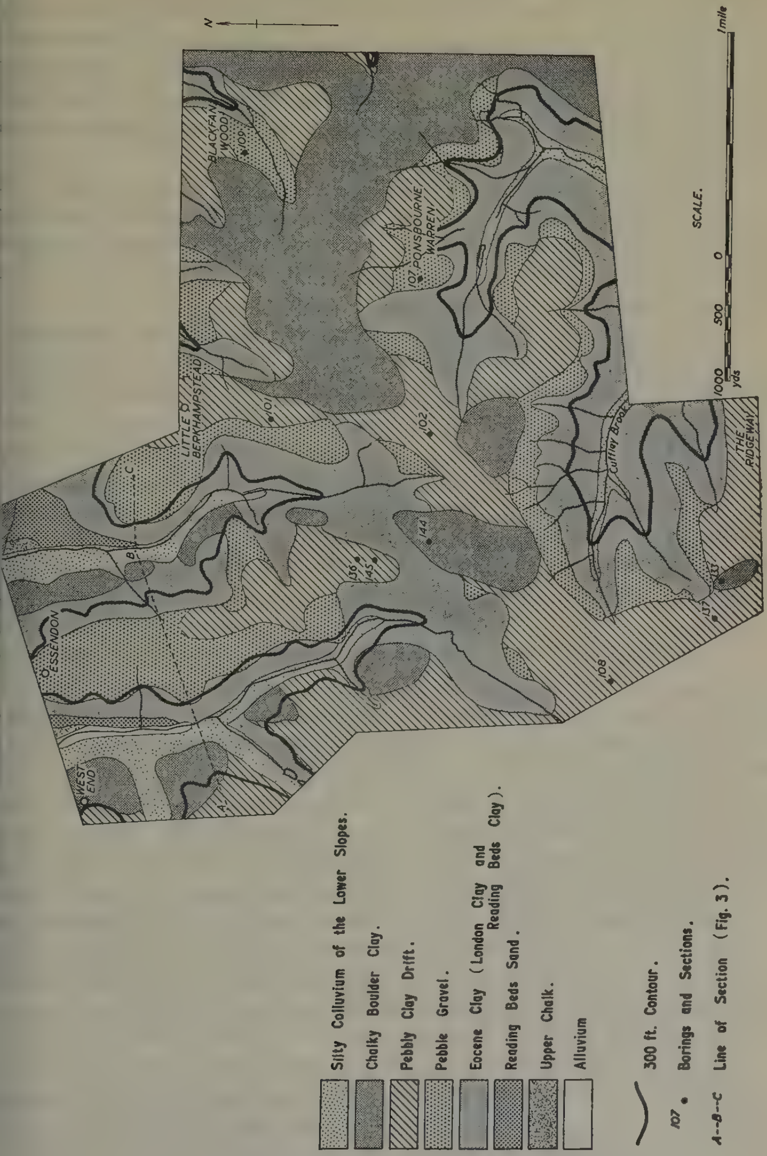

Fig. 1. Drift geology of part of south-east Herts

and bleached coarse sand-grains are conspicuous, giving the impression of a simple outcrop of Pebble Gravel. Below this depth, however, the profile becomes heavier, and sandy clay or sandy clay loam textures are common. The colour may be dominantly grey, or grey strongly mottled with brown or reddish colours. This confused layer usually rests on brown weathering London Clay but further investigation has shown that this is often penetrated by thin bands of coarse Pebble Gravel material.

It is difficult to avoid the conclusion that weathering and eluviation have produced sandy surface layers and heavier-textured subsoils from an originally heterogeneous solifluction deposit on these sloping sites. Only in a few favoured sites are soils formed directly on Pebble Gravel *in situ* without obvious superficial additions of other drifts.

A typical section of Pebble Gravel recently visited by the Associatio occurs in Blackfan Wood (52/301073, elevation 350 feet) showing abou five feet of gravel with a matrix of pale-coloured coarse sand. The assemblage of stones includes Eocene flint pebbles, small white quartz pebbles and subangular flints, with subsidiary amounts of chert from the Lower Greensand, sarsen, and a very few far-travelled stones, mainly quartzites At this site the upper layers of the deposit could easily be contaminated with stones from the nearby Chalky Boulder Clay. The section at Ponsbourne Warren (300062, elevation 350 feet) is of a different type. Mechanical analysis showed that the upper twelve inches of brown sandy loam contained 10% clay. From twelve to twenty-nine inches an orange sandy clay loam strongly mottled with grey and reddish colours contained 34% clay. Below twenty-nine inches reddish coloured sandy loam with 19% clay continued down to eight feet. This material is weakly bedded, with occasional disturbed bands of flint pebbles, quartz pebbles and subangula flint. Channels of grey clay, about a half inch thick, run vertically and horizontally through the sandy loam. No far-travelled stones were found at this site. This section poses a number of pedological problems. Silt (grain size 0.002–0.05 mm.) contributes 16% of the sample in the upper twelve inches, but is virtually absent below that depth. This, along with the other features, suggests that the deposit has suffered a great deal of alteration by weathering subsequent to its deposition. Both the sand mineralogy, and evidence from other sections, indicate that the sandy beds and the coarse gravels are intimately associated and probably two phases of a fluviatile deposit.

(*iii*) *Pebbly Clay Drift*. This occupies much of the level or gently sloping terrain above 350 feet. The soils developed on it are strongly acid under natural conditions, are silty or loamy rather than sandy in texture, and are poorly drained. The clay substratum is not sandy, although thin bands of sand, or sandy-clay textures, occur sporadically at depth as the Pebble Gravel is approached. The deposit is moderately stony throughout and pebbles of Bunter Quartzite are common at the surface. A temporary section at 'The Ridgeway' (291035, elevation 370 feet) showed ten feet of unstratified pebbly clay. Brown, ochreous and grey mottlings were prominent in the upper part, but much reddish mottling appeared below three feet. The stones were mainly pale or blue flint pebbles, with many subangular flints, and in the smaller grades, half to one inch in diameter, numerous pebbles and subangular fragments of red quartzite, sandstone and vein quartz. White quartz pebbles occurred at all levels in subsidiary amounts. Farther east at a lower level the base of the section consisted of two feet of reddish gravel resting on brown weathering London Clay. Still farther east, the pebbly clay became thinner, changing gradually at depth to clayey sand with only occasional pebbles, and finally to coarse pale-coloured gravel similar to that in the Blackfan Wood section. No typical Bunter quartzite pebbles were found below three feet in these sections, but the small red quartzite and sandstone pebbles were recovered from about seven feet, a depth which makes it unlikely that they were introduced by solifluction from later deposits.

The stone assemblage appeared to be richer in far-travelled stones than neighbouring deposits of Pebble Gravel. This conclusion must, however, be tentative owing to the poor quality of most sections and the difficulty of distinguishing sarsen from varieties of quartzite. On its general morphology and texture alone, there seems to be good reason to separate this Pebbly Clay Drift from the underlying Pebble Gravel.

A number of deeper borings were put down into the Pebbly Clay Drift. Borings 101 and 102 (see Fig. 1) passed through brightly mottled clay to a depth of nine feet. In 108 and 137 a sandy gravelly layer assumed to be the Pebble Gravel was reached at seven feet, and water rose rapidly in the hole. In boring 145 flecks of secondary calcium carbonate appeared below five feet in an otherwise non-calcareous brightly mottled clay; no hard chalk was found. At seven feet stoneless dark brown and grey weathered London Clay was reached. The nearby boring 136 passed through similar mottled clay with secondary carbonate before reaching Pebble Gravel. As these two borings were not typical and situated slightly below the plateau surface, there is a possibility that carbonate was deposited from lime-rich water moving laterally over the junction with the London Clay.

(*iv*) *Chalky Boulder Clay*. This occupies similar plateau sites to the previous deposit, and on level ground the boundary between the two can only be mapped by examination of the soil profile. Bunter Quartzite and red sandstone pebbles appear with equal frequency at the surface on both types of drift. The plough layer is normally of clay loam or finer texture and somewhat less stony than on the previous deposit. The immediate subsoil is often similar to the Pebbly Clay Drift, but at eighteen to twenty-four inches the clay becomes a brownish olive colour, and small fragments of hard chalk appear. Shell fragments and striated, faceted lumps of chalk were recovered from the subsoil. Below three feet the weathered till is dominantly grey mottled with brown.

In an attempt to discover the relationship between the Chalky Boulder Clay at 350–400 feet on plateau sites, and the main mass of similar material at lower levels between Hatfield and Hertford, the intervening ground was mapped in detail. Patches of Chalky Boulder Clay were found occupying east-facing slopes of the valleys on either side of Essendon, at heights steadily decreasing towards the north, but maintaining a similar elevation above the valley bottom. Around West End and Essendon the till merges with the main mass of glacial deposits at 250 feet overlooking the recently entrenched Lea Valley.

The distribution of Boulder Clay in these valleys will be enlarged on later, but the pattern is consonant with the view that these patches represent relics of a more extensive fill of boulder clay which has been removed from the steeper west-facing slopes, and from the east-facing slopes where tributary streams join the main valleys. Throughout the sequence the soil profiles developed on this deposit are essentially the same, both in the morphology of the subsoil and in the presence of hard chalk within twenty-four inches of the surface.

(*v*) *Silty Colluvium of the Lower Slopes.* This material consists of a plough layer of silt loam or silty clay loam texture. The immediate subsoil is normally a brown silty clay, often unmottled where the deposit is deep. At its upslope margins it is frequently stony, but becomes less stony downslope. The conspicuous feature of this drift is its asymmetrical distribution in the two north–south valleys near Essendon. In its present position it obviously post-dates the Chalky Boulder Clay, but its origin is obscure.

(b) Mineralogy of the Sands

With a view to obtaining further information on the origin of the Pebbly Clay Drift, mineralogical studies of the sands (.18–.06 mm.) from a number of samples were compared with those from samples of Pebble Gravel and London Clay. The sands were separated in bromoform and the distribution of non-opaque heavy minerals was determined by counts of 150–200 grains from each sample (Table I). The analyses of the Pebble Gravels are comparable with those obtained by Wooldridge (1927, 124) from the same area, and yield suites comprising almost exclusively zircon, tourmaline, staurolite, rutile and kyanite with only minor amounts of other minerals. As epidote, garnet and hornblende occur in appreciable amounts in the Pebbly Clay Drift a source other than the Pebble Gravel is indicated. The amount of feldspar is also greater in the Pebbly Clay Drift, ranging from 7–11% of the light minerals compared with 3–7% in the Pebble Gravel.

The London Clay contained only very small amounts of sand, and did not appear to be a likely source for sands in the Pebbly Clay Drift. The mineralogical assemblage is also very different, opaques constituting over 90% of the heavy separate, while the light minerals are dominated by muscovite and chlorite.

Two samples of Chalky Boulder Clay were also examined. Sample 95 was obtained from near Welwyn Garden City, outside the area studied in Fig. 1. Both samples showed the presence of fairly abundant hornblende and garnet, with epidote conspicuous, an association which appears to be typical of Chalky Boulder Clays in Eastern England (Solomon, 1932; Solomon *in* Clayton, 1957).

TABLE I. **Percentage Distribution of Non-opaque Heavy Minerals**

	Pebbly Clay Drift			Pebble Gravel			Chalky Boulder Clay	
Sample	101	145	108A	108B	109	107	144	95
Zircon	36	22	35	42	38	53	26	26
Tourmaline	23	31	29	36	28	35	18	8
Staurolite	9	12	11	11	12	5	6	3
Garnet	7	11	3	—	—	—	17	12
Rutile	8	4	8	3	8	5	2	2
Apatite	—	—	—	—	—	—	6	2
Epidote	8	9	5	1	—	—	6	6
Hornblende	3	2	3	—	—	—	9	36
Collophane	—	—	—	—	—	—	2	3
Kyanite	3	7	3	4	12	1	2	2
Tremolite	—	1	2	—	—	—	1	—
Monazite	—	—	1	1	—	—	1	—
Chlorite	1	—	—	—	—	—	1	—
Biotite	—	1	—	—	—	—	2	—
Others	2	2	—	2	2	1	1	1

KEY TO SAMPLES

Sample No.	Location	Grid Ref. 52 (TL)	Elevation feet	Depth feet
101	Epping House	292072	375	7
145	See Map	281066	365	4
108A	See Map	271047	400	4
108B	See Map	271047	400	7
109	Blackfan Wood	301073	350	4
107	Ponsbourne Warren	300062	350	6
144	See Map	283061	400	3
95	Holwell Court	270105	250	4

As the minerals differentiating the Pebbly Clay Drift from the Pebble Gravel are also conspicuous in the Chalky Boulder Clay, the possibility arises that the Chalky Boulder Clay has contributed materials to the Pebbly Clay Drift of which the more easily weatherable minerals such as apatite and collophane have been lost during subsequent weathering. This view cannot be entirely refuted, but the apparent absence of large Bunter Quartzite pebbles at depth in the Pebbly Clay Drift, whereas these stones are ubiquitous within the Chalky Boulder Clay and over the general surface of the plateau, suggests that such contamination could only be superficial. This leads to a consideration of the more likely explanation that the garnet, epidote and hornblende suite of minerals is derived in both Pebbly Clay Drift and Chalky Boulder Clay from sources outside the London Basin.

(c) Discussion and Conclusions

The previous literature on this area has dealt mainly with the Pebble Gravel. McKenny-Hughes (1868), Prestwich (1890) and Barrow (1919) attributed this gravel to a Pliocene marine transgression. These writers were mainly impressed by the presence of this uniform flint and quartzose gravel occupying many of the hilltops at 400–500 feet in south Herts and south Bucks. Sherlock (1924a and b) considered it to be the remains of the Barton Beds and other Eocene strata mixed and spread out by local ice and snow. He did not agree that there was any accordance of levels in the distribution of the gravel and mentioned that often the pebbles were in a clay matrix. Wooldridge (1927) and Wooldridge & Linton (1955) divided the Pebble Gravel into two stages on a basis of elevation and lithology, of which the deposits in this area fall into the later fluviatile phase. Attention was drawn to the presence of chert from the Lower Greensand within the gravel, and it was suggested that this material was brought by southern tributaries of an early Proto-Thames flowing approximately along the line of the Vale of St. Albans at a level of about 350 feet.

As the present study has attempted to show, the Pebble Gravel is extremely difficult to map. Frequently where it appears to be best represented at the surface, it is demonstrably heterogeneous and not in its original position. Nor is it a continuous layer under the later drifts. Borings 133, 145 and 144 (see Fig. 1) all passed into London Clay without encountering the Pebble Gravel. Even in the apparently undisturbed sections there is considerable variation in lithology, as is shown by the contrast between the sections in Ponsbourne Warren and Blackfan Wood. The general impression is of an ancient fluviatile deposit much disturbed by solifluction, denudation and glacial erosion. The part played by weathering and soil-forming processes is difficult to assess. In Ponsbourne Warren, the texture of the underlying material consisting almost entirely of sand and clay with virtually no silt is difficult to explain as a purely sedimentary deposit, also the pale colour of much of the gravel in close proximity to the impermeable London Clay, as at Blackfan Wood and The Ridgeway, may be a result of loss of sesquioxides through lateral movement of seepage water rather than the original form in which the material was deposited. The Pebble Gravel in this area can only be roughly dated by its relationship to other deposits, and, in the present state of our knowledge, an early Pleistocene date seems likely.

The Chalky Boulder Clays in the Vale of St. Albans and on the plateau of SE. Herts appear to be similar in lithology, sand mineralogy and in the degree of weathering as shown by the depth of leaching of carbonates. Patches of Chalky Boulder Clay associated with east-facing slopes occur at all elevations from 400 feet down to the 250 feet level near the River Lea. It therefore seems reasonable to conclude that all the obviously chalky till in this area is of similar age irrespective of elevation.

Both the stone content and the sand mineralogy of the Pebbly Clay Drift suggest that it contains materials which could not have been derived from a purely local mixture of London Clay and Pebble Gravel. It is generally leached free of Chalk, and although secondary carbonate was observed in two boreholes, eleven feet of non-calcareous drift was seen at The Ridgeway. These features, along with the prominent red mottling, suggest that it is either of different origin from the Chalky Boulder Clay, or has been subjected to a much longer period of weathering. Without adequate sections it is difficult to be precise about its origins, but if we accept the interpretation of the Pebble Gravel as the deposit of north-flowing tributaries of a proto-Thames following the line of the Vale of St. Albans (Wooldridge & Linton, 1955), then the relief conditions are unsuitable for the deposition of the Pebbly Clay as a 'Head' by solifluction from the north or west. The only remaining alternative is to consider this drift as a weathered boulder clay. No clear case of Chalky Boulder Clay superimposed on Pebbly Clay Drift can be produced, but a comparison of the distribution of these deposits in Fig. 1 suggests that the Chalky Boulder Clay ice advanced into a moderately dissected terrain. Where the Pebbly Clay is found at comparatively low level it appears to have suffered slumping down-slope, like the Pebble Gravel, by solifluction, and is directly associated with a thicker source material upslope. This suggests that it was laid down on a fairly level undissected terrain which pre-dates the onset of the Chalky Boulder Clay glaciation.

While this interpretation satisfies the local conditions, it must be emphasised that the dating of Boulder Clays by a comparison of topographical position and degree of weathering as shown by the leaching of carbonates is susceptible to a number of errors. Tills of similar age may vary considerably in chalk content and texture, which will obviously affect the ease with which decalcification takes place. This process is also likely to be more rapid on level sites than on sloping sites where much of the rainfall may run off as surface water. In addition, accessions of non-calcareous solifluction or wind-blown materials over the till require careful study to avoid confusion with decalcified till.

The sequence in this restricted area may be summarised as follows:

(5) Further denudation.
(4) Chalky Boulder Clay;
(3) Dissection, weathering and denudation;
(2) Pebbly Clay Drift;
(1) Pebble Gravel;

The correlation of such a scheme with neighbouring areas raises some problems. In a study of glacial deposits around Chelmsford, Clayton (1957) recognised the deposits of two distinct glaciations, the later one consisting of a tripartite sequence of two tills separated by gravels which he called Maldon Till, Chelmsford Gravels and Springfield Till, and equated with the Saale glaciation of North Germany. Where either of these tills occurred at the surface, clay with chalk fragments was found within a depth of about two feet. On higher ground to the south of these deposits an older till could be distinguished, which appeared to be separated from the earlier group by an erosional break of about 100 feet. This was named the Hanningfield Till and characterised by the absence of chalk in the upper weathered zone. Chalk-rich till was, however, found at many sites between six and fourteen feet below the surface. In the Harlow area a similar topographical sequence was found, except that the Hanningfield Till, although occupying the elevated surface of Epping Long Green at 350 feet, was nowhere leached free of chalk to any great depth. This was attributed to a loss of the leached zone by erosion. As Clayton suggests, the tripartite sequence can be extended westwards into Hertfordshire, around Hertford and the Vale of St. Albans. The elevated tracts of SE. Herts are, however, partly occupied by Chalky Boulder Clay of Maldon or Springfield type, and a correlation of the Pebbly Clay Drift with the Hanningfield Till is uncertain, as no unweathered calcareous lower zone of this drift has been found in the course of the survey.

In view of the similarity of the terrain in SE. Herts and the Harlow area, and the fact that the two localities are only fifteen miles apart, it is pertinent to suggest that the higher chalky till on Epping Long Green is not necessarily of greater age than the Maldon and Springfield Tills at lower level. Without some evidence of deep weathering or a different suite of constituent materials, topographical evidence alone would appear to be insufficient grounds for the separation of two neighbouring Chalky Boulder Clays.

Wooldridge & Linton (1955) have drawn attention to the belt of drift between Welwyn and Beaconsfield within which are weathered clayey deposits with far-travelled stones. These lie mainly around 350–400 feet along the northern edge of the Vale of St. Albans, and were named the Chiltern Drift; interpreted as an ancient till. This terrain has not been studied in detail by the Soil Survey and the only objection to correlating this material with the Pebbly Clay Drift is that its position along the Chiltern dipslope plateau raises the possibility of an origin by solifluction. If, however, both drifts can be considered as true boulder clays, then they must be the ground moraine of ice which passed round the Chilterns through the Goring or Stevenage gaps, as the higher parts of the Chilterns are conspicuously free of far-travelled stones.

The limited area studied here by soil survey methods has led to a better appreciation of the nature of the drift deposits, but because of the small extent of this survey, an extension of the local sequence into the general field of Pleistocene chronology would appear premature until more detailed work has been carried out in the areas between this district and the classic sequences of East Anglia and Thames Valley.

3. ASYMMETRIC VALLEYS

(a) Description of the Terrain

South-east Hertfordshire and neighbouring parts of north Middlesex are areas of generally subdued relief. The relief amplitude is seldom more than 150 feet, and from plateau or divide to the valley bottoms is usually of the order of 100 feet. In spite of this, the differing steepness of slopes having different aspects can be distinguished quite readily as shown in Fig. 2. Here the asymmetric reaches of valleys are shown and the steeper slope indicated. The map was drawn from a study of the 1 : 25,000 maps, with additional field work in doubtful cases. Only the more conspicuously asymmetric reaches of the valleys are included. A point which stands out from the map is that the middle reaches of valleys aligned north–south and north-west–south-east are mainly affected, whereas east–west valleys are normally symmetrical. The differences in angle of slope are 8–10° for the steeper slopes and 5–6° for gentler slopes.

A detailed cross-section of two of these valleys is shown in Fig. 3, along the line ABC in Fig. 1. The steeper west-facing slopes are free of drift, and soils are formed directly on Eocene Clay, with the exception of a narrow outcrop of Reading Beds Sand. On the gentler east-facing slope Chalky Boulder Clay occurs above a footslope mantled with silty colluvium. There is little likelihood that the Chalky Boulder Clay could have slipped into its present position by solifluction, for no obvious source area appears upslope from either of these patches, and in addition Chalky Boulder Clay would quickly lose its identity if subjected to down-slope movement in an

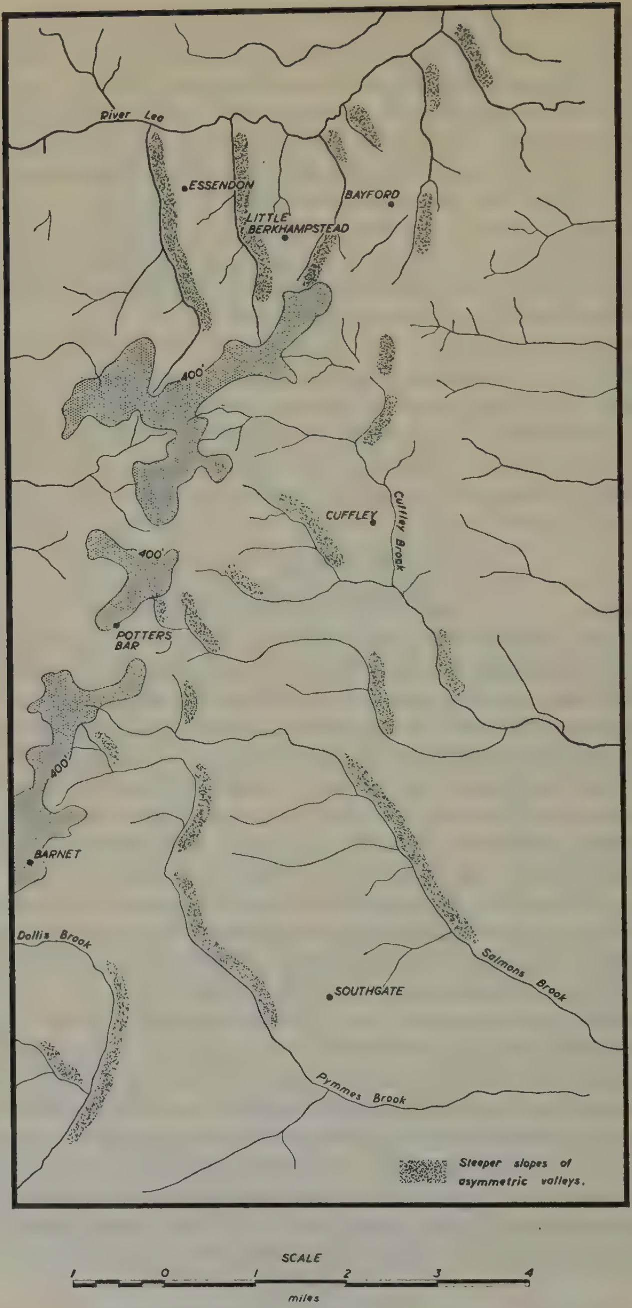

Fig. 2. Asymmetric valleys of south-east Herts and north Middlesex

area of dominantly non-calcareous drifts such as this. We must therefore conclude that the Chalky Boulder Clay is in the position where it was originally deposited as ground moraine, and also, more surprisingly, the slope on which it rests can have suffered very little change in form since the time of its deposition. No borings succeeded in reaching the base of the till on this slope, but as it does not occur in the tributary valleys its depth is unlikely to be more than about fifteen feet.

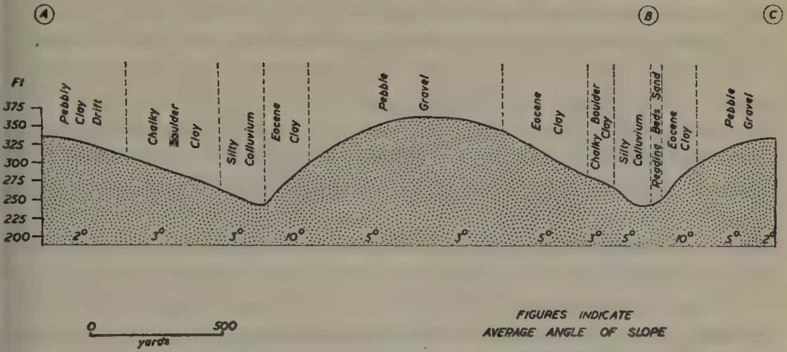

Fig. 3. Cross-section of valleys near Essendon

Before considering the ultimate cause of asymmetric valleys it is necessary to discuss the sequence of events indicated by this pattern of relief and drift deposits in Fig. 3. It is difficult to determine whether the original valley before the onset of the Chalky Boulder Clay ice was symmetric or otherwise, but the valley bottom was probably somewhere near the lower margin of the till. Support for this hypothesis is found in the slight 'shelf' feature discernible on the ground at this level at many sites, and in addition a thin belt of gravel, too narrow to appear on the map, often marks the lower edge of the till, causing a general stoniness in the neighbouring silty colluvium. This gravel may be fluvio-glacial or a simple stream-laid deposit, but it inevitably suggests an earlier graded reach, well above the present valley bottom. It seems likely that both slopes of these relatively shallow valleys were covered with a layer of Chalky Boulder Clay when the ice sheet depositing this material extended on to the plateau to the south. Denudation after this period has deepened the valleys by a further twenty-five feet and caused the stream to shift laterally towards the east. There can be little doubt of the efficacy of this lateral movement after the deposition of the Chalky Boulder Clay for the lower reaches of both valleys farther north are cut entirely in glacial deposits and are markedly asymmetric (Fig. 2). The proposal that ice moving from the east has denuded and steepened the west-facing slope is therefore untenable.

(b) Discussion and Conclusions

There is much evidence to show that asymmetric valleys are of widespread occurrence in Middle and Arctic latitudes of the northern hemisphere, and a large number of theories have been proposed to explain these phenomena. The literature on this subject was reviewed by Ollier & Thomasson (1957). All writers who have studied this type of valley are agreed that they are *erosional* in form and do not result from a fill of drift in a normal symmetrical valley. Büdel (1944) maintained that these valleys were formed under periglacial conditions when dominant westerly winds deposited snow more thickly on east-facing lee slopes than on the opposing slopes. This slope would retain its snow-cover for longer periods owing partly to the greater depth of snow, and partly to its cooler aspect as it faces the morning sun, which is generally not so effective in raising ground temperatures as the afternoon sun (Geiger, 1950). It was proposed that this excess of unmelted snow would cause strong solifluction on this slope, and reduction—in the Davisian sense—caused it to become less steep. While this theory contains many useful features, it is untenable as an explanation of the valleys in Fig. 3, because any considerable reduction of the east-facing slope would have removed all traces of the Chalky Boulder Clay, and there is conclusive evidence that the period of asymmetric development post-dates the deposition of the till.

The alternative hypothesis proposed by Ollier & Thomasson is that excessive erosion and waste-removal was confined to the west-facing slopes, causing it to become steeper and undergo parallel retreat. In the valley bottom the stream continued down-cutting, but shifted laterally to maintain a position at the foot of the more active slope. On the east-facing slope the processes of denudation operated more slowly and its major form remained unchanged. This theory explains both the preservation of Chalky Boulder Clay on east-facing slopes and its absence from the opposing slopes.

Numerous other theories of the origin of asymmetric valleys can be shown to be inadequate owing to their wide distribution. Clayton (1957) suggested that similar valleys in Essex were caused by tectonic down-warping towards the North Sea. It is by no means certain that such warping would result in steeper west-facing slopes, and a consideration of the widespread occurrence of such valleys in Belgium (Geukens, 1947), Bavaria (Büdel, 1944) and France (Pinchemel, 1954) makes it unlikely that similar tilting or warping has occurred in all these areas.

Structurally controlled uniclinal shifting of streams and rivers cannot be invoked in the case of the London Clay which is a uniform material without any conspicuously hard layers. Finally, the tendency of moving bodies of water to be deflected to the right in the northern hemisphere (cf. ocean currents) is inapplicable as west-facing slopes are generally steeper whether the main stream flows north or south.

Although the results of asymmetric development can be described in detail, the ultimate cause of these phenomena is difficult to ascertain. It is obviously desirable that any hypothesis erected should have widespread applicability, and for this a climatic explanation seems most appropriate. As mentioned earlier, ground temperatures vary considerably with aspect. In general the south-westerly aspect is warmest throughout most of the year, and a westerly aspect is warmer than one facing east. At least one period of periglacial climate has occurred since the deposition of the Chalky Boulder Clay in this region, and under such conditions seasonal or diurnal thawing could occur more frequently on the warmer slopes facing south-west than on similar slopes facing east. This might lead to stronger solifluction, or possibly a continuance of normal water erosion on these sites while other sites were thickly covered with snow. However, insolation need not necessarily only influence the processes of denudation in periglacial climates. The connection between frozen ground and asymmetric valleys has been inferred by many writers because many of these valleys on chalk country and permeable sands and gravels are at present dry. Their widespread occurrence on a normal terrain with surface water such as the London Clay raises the possibility that interglacial climates may be suitable for their formation. Nor is insolation the sole factor involved, as a study of Fig. 2 illustrates. Valleys aligned east–west are normally symmetrical, and although the south-west aspect is warmer than one facing south, both would be considerably warmer than the directly north aspect. For this reason, it may be necessary to reconsider the question of wind direction raised by Büdel (1944). The effect of a deep covering of snow may be to protect the ground rather than accelerate erosion.

The present study does not enable a full explanation of the origins of asymmetric valleys to be constructed. However, some conclusions concerning their formation can be reached. A period of asymmetric development occurred after the deposition of the main Chalky Boulder Clay of this area. The more gentle east-facing slopes suffered relatively little reduction during this process. This type of valley is not confined to permeable deposits, but also occurs on impermeable terrain with normal surface drainage.

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The Detrital Mineralogy of the Cretaceous Rocks of South-East Devon, with Particular Reference to the Cenomanian

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BSTRACT: The sand in the Foxmould was largely derived from pre-Cretaceous ediments such as the New Red Sandstone. This source of sand became much less nportant during the accumulation of the Chert Beds, the small amount of sand in these eds indicating a sparse supply of detritus from the Armorican granitic complexes of ıe Cornubian massif. Increased influxes of granitic detritus direct from this massif ntered south-east Devon in late Albian and early Cenomanian times to give the 'Top andstones' of the Upper Greensand, the Cenomanian sands of Wilmington facies and ossibly Division A1 of the Cenomanian Limestone. The direct supply of granitic etritus became more and more restricted as the Cenomanian sea spread over the 'ornubian massif, the sand in the younger Cenomanian beds, and the Middle Chalk, eing almost entirely derived from early Cenomanian and late Albian deposits which

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underwent erosion in regions of localised tectonic uplift. It is suggested that during Upper Cenomanian times and during the accumulation of the Middle Chalk the Cornubian massif was submerged beneath the sea.

1. INTRODUCTION

(a) Summary of the Stratigraphy

THE OLDEST Cretaceous rocks of south-east Devon comprise an extensive highly dissected, gently dipping sheet of glauconitic sands and calcareous sediments, of Middle and Upper Albian age, and known collectively as the Upper Greensand. To the east of Seaton the basal beds are sufficiently argillaceous to justify the use of the name 'Gault' but over most of the area

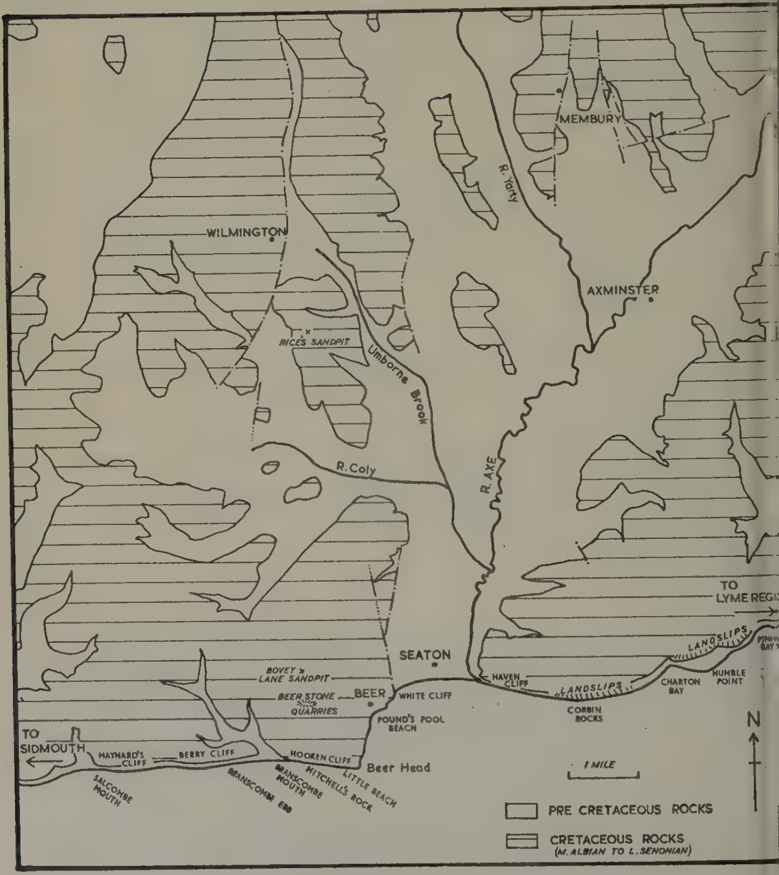


Fig. 1. Geological sketch-map of south-east Devon showing position of localities referred to in the text

the basal beds are sandier and their separation as an argillaceous formation distinct from the Upper Greensand is impracticable. The base of the Upper Greensand is strongly transgressive. East of Membury, Axminster and Culverhole the Cretaceous rocks rest unconformably upon easterly dipping Lias, farther west they overstep on to the Keuper Marl and in the two small outliers forming the Haldon Hills, north of Teignmouth, they rest directly upon New Red Sandstone breccias.

The Cenomanian and Turonian rocks are preserved as outliers, the largest of which extends for about two miles north of the coast between Lyme Regis and Seaton (Smith, 1961). Another important outlier extends for a similar distance northwards from the coast between Seaton Hole and Branscombe Mouth, its eastern margin being determined by the Seaton Hole Fault. Cenomanian and Turonian deposits also cap the high ground terminating in the cliffs west of Branscombe Mouth, the westernmost development being that at Maynard's Cliff, east of Sidmouth. Farther inland, a faulted chalk outlier occurs in the Wilmington district, about seven miles north of Branscombe Mouth, and another is situated in the Membury district, close to the Devon–Somerset border.

(*i*) *The Upper Greensand* (Fig. 2). The Upper Greensand in south-east Devon is about 150–200 feet thick, the thickness decreasing from east to west. It is conventionally subdivided into a lower group of somewhat argillaceous, glauconitic sands termed the 'Foxmould', and an upper group of calcareous sandy deposits, the 'Chert Beds', in which layers of chert-nodules are irregularly distributed. The uppermost eight to thirteen feet of the Upper Greensand rarely contains chert, apart from occasional transgressive veins. It is distinctly sandier than the beds below and often contains conspicuous wavy streaks of greensand. Towards the west it passes into consolidated shell-sands, well exposed at Kempstone Rocks, Dunscombe. This uppermost division of the Upper Greensand will be termed the 'Top Sandstones'. To the west of Seaton Hole the base of the Top Sandstones is defined by a thin but persistent bed of pebbly greensand which forms a conspicuous recess in cliff-sections and which will be referred to as the 'Coarse Band'. The Coarse Band thins out eastwards and is absent from the sections east of the River Axe. In these sections there is no clearly defined base to the Top Sandstones although the highest layer of chert-nodules provides a rough indication. The upper surface of the Top Sandstones is a well-defined erosion-surface throughout south-east Devon.

(*ii*) *The Cenomanian Deposits*. In the Membury district, on the borders of Devon and Somerset, the Cenomanian is represented by about fifty feet of Lower Chalk possessing a well-defined glauconitic basement-bed, rich in derived or *remanié* Lower and Middle Cenomanian fossils and phosphatised pebbles of limestone (Hancock, *in* Smith, 1957b). In the coastal outliers between Lyme Regis and Sidmouth there are no such deposits of

Lower Chalk, the Cenomanian being represented by a few feet of indurated sandy limestones collectively known as the Cenomanian Limestone. The Cenomanian Limestone is mainly of Lower Cenomanian age and consists of four beds which are conveniently referred to as Divisions A1, A2, B and C (Fig. 2 and Smith, 1961). The relative thicknesses of these beds vary greatly from place to place, indeed one or more are usually absent at any one locality. Sections displaying the superposition of all four divisions occur only where the formation as a whole is exceptionally thick, as in the Hooken Landslip (Smith, 1961).

In the Wilmington outlier there is also no Lower Chalk between the Upper Greensand and Middle Chalk. In this outlier the Middle Chalk rests non-sequentially upon Lower Cenomanian deposits comprising a variable, but often considerable, thickness of friable calcareous sands capped by hard calcareous sandstone ('grizzle') and a thin limestone resembling Division B of the Cenomanian Limestone (Smith, 1957b). Similar deposits also occur in the Beer district (Smith, 1961), and will be referred to as 'Cenomanian sands of Wilmington facies'.

(*iii*) *The Middle and Upper Chalk*. The White Chalk of south-east Devon has been described by Rowe (1903) and Jukes-Browne (1903, 1904). It has not been studied in detail in the present work since it normally contains very scarce sand-grains. The *Inoceramus labiatus* Zone, however, especially the basal part, usually contains scattered grains of quartz, glauconite and other detrital minerals which have been examined qualitatively.

(b) Summary of Previous Work

Very little information has been published concerning the nature and distribution of the detrital minerals in the Cretaceous deposits of south-east Devon. Boswell (1923) restricted his study of the detrital grains in the Cretaceous deposits of south-west England to the Upper Greensand of the Haldon Hills, which are situated beyond the western boundary of the area studied by the present author. Groves (1931) listed the heavy minerals obtained from six samples of Upper Greensand which were collected from widely scattered localities between the Haldon Hills and Lyme Regis. He did not, however, indicate the exact horizons at which the samples were collected and did not state whether they were from the Chert Beds or the underlying Foxmould. Furthermore, he expressed the relative frequencies of the heavy minerals in terms of an arbitrary numerical scale and not as percentages of grain-counts.

Hill (*in* Jukes-Browne, 1903, 274–6) described the acid-insoluble residues from two samples of Cenomanian Limestone from the Beer district but gave no quantitative information concerning the relative frequencies of the detrital minerals contained in them. Groves (1931) gave no fresh information concerning the detrital minerals in the Cenomanian Limestone but

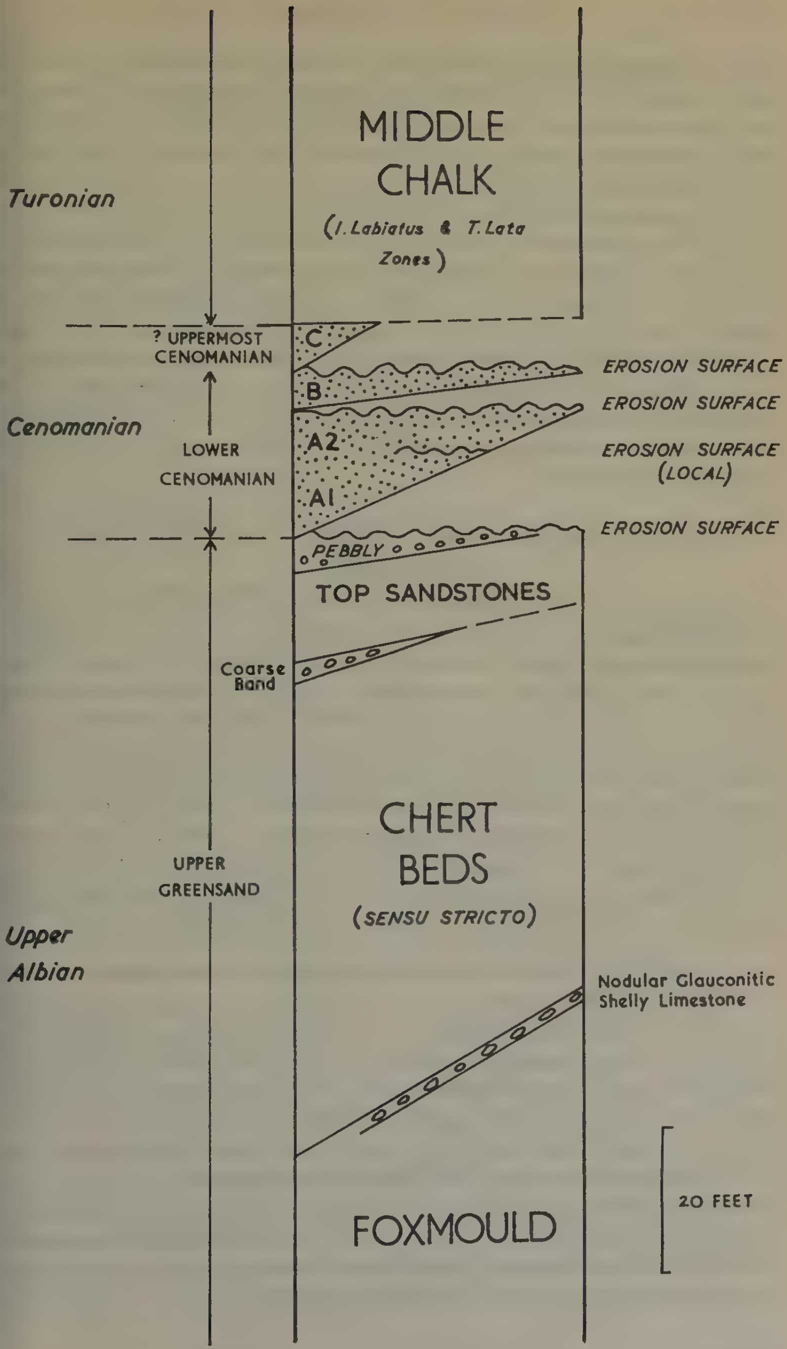


Fig. 2. Diagrammatic section showing the subdivisions of the Upper Greensand, Cenomanian Limestone and Middle Chalk used in the present paper

noted the presence of angular tourmaline and euhedral zircon of Dartmoor type in a sample of Lower Chalk from Lulworth Cove in Dorset. Lists of heavy minerals obtained from the Middle Chalk of south-east Devon have been published by Hume (*in* Jukes-Browne, 1903, 507–10), Macdonald Davies (1919) and Groves (1931, 77).

The foregoing survey of previous work on the detrital minerals in the Cenomanian Limestone and contiguous deposits shows that no systematic quantitative description of their nature and distribution has so far been published. The present paper is intended to remedy this deficiency and is based on the study of more than one hundred acid-insoluble residues, the majority of which were obtained from the various Cenomanian deposits and the formations immediately above and below them. Samples were taken from all the well-exposed coastal sections in these rocks between Sidmouth and Lyme Regis and from inland sections as far north as the Wilmington and Membury outliers.

2. THE INSOLUBLE SANDY RESIDUES OF THE SAMPLES

(a) Sampling

After removing the outermost two or three inches of rock a vertical sequence of samples was collected from each bed and combined to give a single sample considered to be representative of the bed as a whole. The samples were not truly random because it was considered desirable to take them from that part of each exposure displaying the least weathering and which was reasonably free from fissures and encrusting organic growths. Because of the small thickness of most of the Cenomanian beds the vertical sequence of samples usually included the entire thickness of the bed. In the case of beds more than two feet thick, however, a vertical sequence of six samples was taken at constant vertical intervals between the base and top of the bed. On the other hand, very thin beds were sampled laterally at six equally spaced points along the length of the particular exposure.

(b) The Determination of the Sand-Content of the Samples

After crushing into walnut-size pieces, mixing and repeated quartering about one to two hundred grams of each of the combined samples was treated with dilute hydrochloric acid (20%) and boiled until the insoluble residue had lost its original brown colour. The scum, clay and most of the silt fractions were removed from the insoluble residue by careful decantations, the decanted material being examined to ensure that there was no loss of grains above the lower diameter of the sand grades, viz. $\frac{1}{16}$ mm.; the rest of the residue was then dried and weighed. The silt and clay contents were always very small, rarely exceeding 2% of the weight of the original sample and often much less. Tables I A, I B show the amounts of insoluble sandy residue from the more important samples.

TABLE IA. **The Sand-Content of the Cenomanian Limestone and Uppermost Upper Greensand (W/W), i.e. the Percentage of Acid-Insoluble Constituents exceeding $\frac{1}{16}$ mm. in diameter**

Locality	*Division C*	*Division B*	*Division A2*	*Division A1*	*Uppermost Upper Greensand*
Largest beach block below Maynard's Cliff	×	23*	15	32	—
Largest beach block rear of Branscombe Ebb	×	6	17	32	—
Mitchell's Rock (Hooken Landslip)	×	×	×	24	24
Westernmost Pinnacle (Hooken Landslip)	25	2	6	—	—
Beer Head	×	—	—	29	28
Pound's Pool Beach (about half-way between The Hall and rear of beach)	×	5	7	×	—
Beer Beach (West Side); close to small rock fall	×	11	12	×	22
Whitecliff (large block in rock fall)	×	3	6	×	25
Humble Point (large block at eastern end of Charton Day)	21	9	10	27	16

× Bed absent due to intra-Cenomanian erosion.
* Abnormal deposit, see Smith (1961).

TABLE IB. **The Sand-Content of certain other Cenomanian and Upper Albian Deposits (W/W), for comparison with Table IA**

Sands of Wilmington facies	
White Hart Sand-pit, Wilmington	70
Rice's Sand-pit near Sutton Thorn	97 (almost completely decalcified)
Bovey Lane Sand-pit near Beer	75
Chert Beds, below Coarse Band	
King's Hole, Beer	8
Corbin Rocks	3*
Storridge Hill (near Chardstock)	5

* Total insoluble residue (sand, silt, clay) 12.

(c) Analysis of Results

The sand-content of successive beds in the Cenomanian Limestone tends to decrease upwards, except in the case of Division C and the westerly developments of Division B (Maynard's Cliff). Since all the insoluble sandy residues from Division B contain much more glauconite than the lower beds the upward decrease in the frequency of detrital quartz is even greater than is suggested by Table IA.

The sand-content of Division B tends to be greatest close to regions in which Division A is either poorly represented or completely absent. Contrast, for instance, the sample from the west side of Beer Beach with those from Whitecliff and the Pinnacles (Smith, 1957a).

The sand-content of Division B tends to be smallest where the thickness of this bed is greatest; indeed, in the thickest developments, for example that in the Pinnacles, much of the bed is comparatively pure chalk.

The sand-content of Division A2 tends to vary in a similar manner to that of Division B.

The sand-content of Division A1 is much more uniform than that of the higher beds but tends to decrease in an easterly direction.

The sand-content of the uppermost Upper Greensand, the Top Sandstones, is similar to that of the lowest division of the Cenomanian Limestone but much greater than that of the underlying Chert Beds.

The sand-content of the calcareous sands of Wilmington facies is very much greater than that of any part of the Cenomanian Limestone or Upper Greensand, except the Coarse Band. At Rice's pit the sands are highly disturbed and have evidently been almost completely decalcified. At Wilmington and Bovey Lane the high sand-contents may be due to partial decalcification but the homogeneous character of the sands, particularly those exposed in the White Hart sand-pit, Wilmington, suggests that their original content was higher than that of the other Cenomanian and Upper Albian deposits.

Despite the very low content of quartz and glauconite grains the beds overlying the Foxmould have been called 'sandstones' by all previous authors (Jukes-Browne, 1900, 365–6). The arenaceous appearance of the rocks in question is, however, primarily due to the granular nature of the calcitic constituents. The name 'sandstone' can be applied to them as a field-term but they are more accurately classified as calcarenites, the Top Sandstones being sandy (or quartziferous) calcarenites.

(d) The Mechanical Analysis of the Insoluble Sandy Residues

Diagenetic changes such as grain-growth and recrystallisation prevent the determination of the depositional grain-size of most of the calcareous materials which form the dominant constituents of the Cenomanian deposits. Some indication of the effects of current-sorting can, however, be obtained by studying the range and frequency of particle-sizes in the insoluble sandy residues obtained from samples large enough to provide sufficient material for sieve-analysis. Thirty-six samples, weighing up to three thousand grams, and calculated to yield about one hundred grams of sand, were therefore treated with dilute hydrochloric acid and the insoluble sandy residues separated by the procedure outlined above. The sieves used in this study were B.S.S. 8, 16, 30, 60, 120, 240, and were chosen because of their mesh-apertures approximate to the limits of the grade terms used in the Wentworth Grade Scale. Histograms and cumulative curves were plotted for each sandy residue and the median diameter and sorting coefficient calculated for each sample. Fig. 3 shows a selection of the histograms.

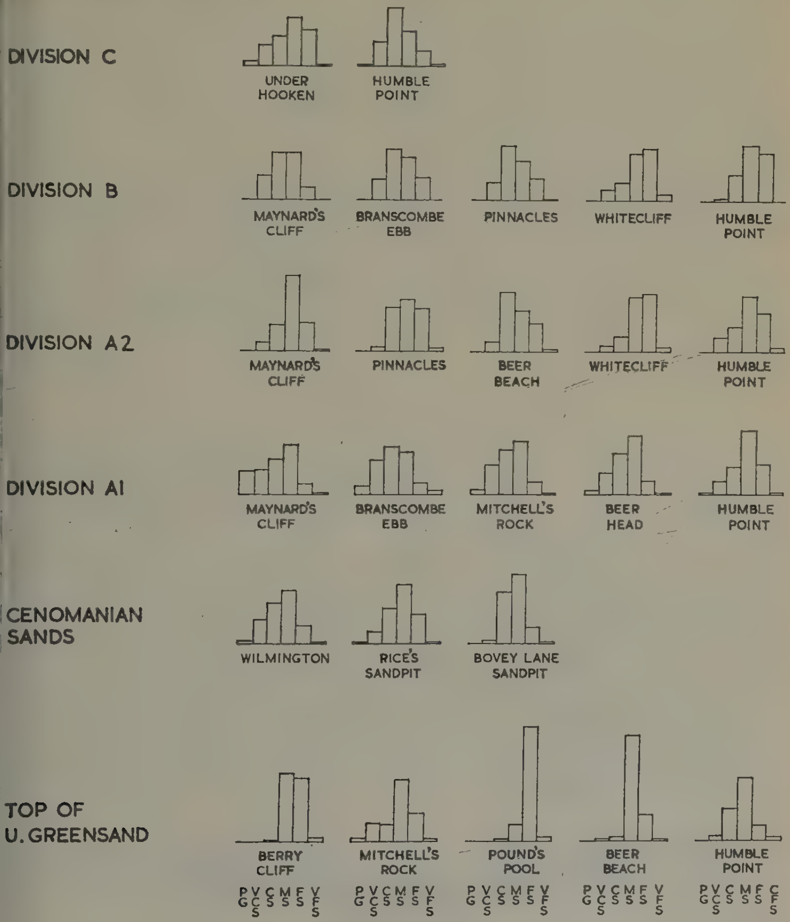


Fig. 3. Histograms of insoluble sandy residues from the Cenomanian deposits and uppermost Upper Greensand of south-east Devon

PG Pebbles and granules VCS Very coarse sand CS Coarse sand
MS Medium sand FS Fine sand VFS Very fine sand

(e) Analysis of Results

The sand in the Top Sandstones is for the most part finer-grained and better sorted than that in the Cenomanian deposits. The uppermost part of the Top Sandstones is, however, locally coarser-grained and less well sorted, in these respects resembling the Coarse Band and the Cenomanian sands. The field-evidence suggests that the presence or absence of these coarse-grained Upper Albian deposits is dependent upon the thickness of Upper Greensand removed during the erosion-period preceding the accumulation of the Cenomanian Limestone. The earliest Cenomanian deposits are likely, therefore, to contain sand derived from the Upper Greensand but the regular easterly grading of the sand in Division A1, and the exceptionally coarse-grained character of the westernmost development of this deposit, suggest that much of the sand originated as a renewed influx of detritus from a similar source to that which provided the sand in the Top Sandstones. The present sporadic occurrences of Division A1 and the Cenomanian sands of Wilmington facies appear to be remnants of a once continuous spread of these early Cenomanian sediments separated as a result of intra-Cenomanian erosion (Smith, 1957a). Although the grading of the sand in Divisions A2, B and C show trends similar to that in Division A1 the lateral changes are much less regular. In Divisions B and C these irregularities are partly due to the irregular distribution of glauconite granules which occur mainly in the medium and fine sand-grades. In all three divisions, however, and in the *Inoceramus labiatus* Zone of the Middle Chalk, coarse grains are most frequent close to areas in which Division A1 is either very thin or absent, and where the Cenomanian Limestone as a whole is greatly attenuated. Taken in conjunction with the evidence provided by the irregular lateral variations in the sand-content, this suggests that during Cenomanian times there was only one major influx of sand from outside south-east Devon, i.e. that found in the Cenomanian sands and Division A1. Most, if not all, of the sand in the younger Cenomanian beds and Middle Chalk was derived from these early Cenomanian deposits, particularly from regions of localised tectonic uplift such as Beer Beach and Branscombe Mouth (Smith, 1957a, 1961). Indeed, sand was probably passed from one bed to the next without the influx of fresh detritus from outside south-east Devon.

3. THE PETROGRAPHY OF THE INSOLUBLE SANDY RESIDUES OF THE CENOMANIAN DEPOSITS

(a) Procedure

Each of the insoluble sandy residues obtained in the determination of the sand-contents was separated into a light and heavy fraction by flotation in bromoform. The light fractions were spread out on black cardboard and examined under a binocular microscope. Permanent mounts of representative grains obtained by repeated quartering were also examined under a petrological microscope. The heavy fractions, reduced where necessary by a single quartering, were mounted in Canada balsam and the relative proportions of the various constituents determined by counting all the grains seen in the successive fields encountered during traverses across the entire mount with a mechanical stage (Tables IIA, IIB). Because of the overwhelming abundance of tourmaline in most of the heavy fractions and the failure of attempts to remove this mineral quantitatively by electromagnetic methods, it was considered desirable to count the maximum number of grains that could be conveniently mounted under a microscope slide cover-glass. In the coarse-grained residues this number was limited by the very large size of some of the tourmaline grains but with few exceptions at least one thousand grains were counted in each mount. Even so, the total number of many of the accessory minerals counted was very small so that the percentages of the less common minerals given in Tables IIA, IIB must be considered as somewhat approximate. In most cases, however, these low percentages serve to demonstrate the absence of major lateral variations in frequency and are therefore sufficiently accurate for the present study.

(b) Heavy Fractions—Opaque Minerals

In most cases about 20% of each heavy fraction consists of opaque minerals. Reddish and brownish particles and mossy aggregates of 'limonite' are abundant in all the Cenomanian deposits, even in the hard, compact limestones in which the mineral grains are sealed in an apparently impermeable calcareous mosaic. Most of this 'limonite' dissolved readily in boiling dilute hydrochloric acid, but some, presumably cryptocrystalline haematite, remained undissolved, even after prolonged boiling. The 'limonite' must be classed as predominantly, if not entirely, authigenic, but except for occasional undoubted pseudomorphs after glauconite, its source is obscure. A considerable amount must have been produced during comparatively recent weathering but some was undoubtedly formed from unstable detrital minerals soon after deposition, before diagenesis led to the sand grains being sealed in an impermeable calcareous matrix. Ilmenite, often coated with leucoxene and/or specks of 'limonite', is invariably present in the heavy fractions, normally as a minor constituent, but magnetite is either completely absent or extremely rare. The absence of magnetite is somewhat surprising but it is noteworthy that Boswell (1923) records it as being absent from the Upper Greensand of the Haldon Hills.

Other opaque minerals which were positively identified include basal sections of tourmaline displaying very strong absorption, andalusite crowded with black inclusions, and adamantine cassiterite. Considerable difficulty was encountered, however, in identifying many of the opaque grains in permanent mounts. No attempt was therefore made to express the various opaque minerals as percentages of grain-counts. Furthermore, it was considered desirable to express the relative abundance of the various transparent minerals in terms of percentages of the total non-opaque grains, the vast majority of which were positively identified (Table II A).

TABLE II A. **The Principal Transparent Heavy Minerals in the Cenomanian Deposits of South-East Devonshire Expressed as Percentages of the Total Transparent Grains**

SAMPLE	*Tt	Tb	Tp	An	Zi	St	Ru	N
Cenomanian Limestone								
Division C								
Pinnacles	92	2.4	2.1	1.0	2.7	1.6	0.4	1508
Humble Point	77	2.8	1.1	0.3	19.8	0.5	1.7	1261
Division B								
Maynard's Cliff	92	1.9	2.6	1.5	2.6	0.7	0.4	1342
Branscombe Ebb	86	2.9	2.9	2.0	7.9	0.1	0.9	836
Pinnacles	94	5.5	1.3	1.0	2.3	0.6	0.6	1500
Beer Beach (West Side)	95	5.3	1.7	0.4	0.8	1.3	0.8	1172
Whitecliff	88	2.5	2.1	0.9	8.2	0.5	0.7	1183
Humble Point	83	1.7	1.0	0.5	12.4	0.7	0.8	1169
Division A2								
Maynard's Cliff	92	2.4	1.8	0.4	2.4	0.4	0.9	935
Pinnacles	85	6.3	2.1	0.4	4.2	1.3	0.8	1332
Beer Beach (West Side)	83	1.9	1.3	1.1	10.2	1.0	2.2	1746
Whitecliff	71	1.8	1.8	0.2	22.3	0.9	2.6	2135
Humble Point	81	1.6	3.1	0.5	9.6	2.4	1.5	1662
Division A1								
Maynard's Cliff	89	0.4	4.5	1.9	2.2	1.5	0.4	591
Branscombe Ebb	80	3.7	2.9	0.7	9.7	1.2	1.5	882
Mitchell's Rock (Hooken Landslip)	82	3.2	3.0	0.2	4.8	4.2	2.0	801
Beer Head	87	2.8	3.7	0.6	4.5	1.1	0.5	920
Humble Point	81	3.3	1.5	0.4	7.9	3.1	2.8	1475
Cenomanian Sands of Wilmington facies								
White Hart (Wilmington) sand-pit	90	4.5	2.4	1.3	2.8	2.6	0.7	808
Bovey Lane sand-pit, Beer	93	4.8	2.5	1.3	2.2	0.9	0.4	1171

* Explanation of symbols: Tt: Total tourmaline, Tb: Blue and parti-coloured tourmaline, Tp: Topaz, An: Andalusite, Zi: Zircon, St: Staurolite, Ru: Rutile, N: Total number of grains counted.

(c) Heavy Fractions—Transparent Minerals

(*i*) *Tourmaline*. All the heavy fractions obtained from the Cenomanian deposits of south-east Devon contain a flood of tourmaline grains displaying a wide variety in size and shape (Table II A, Plate 12 A). The majority of the grains are subangular to sub-rounded fragments with diameters ranging from 0.1 to 0.5 mm. Larger ovoid or spheroidal grains and granules with diameters exceeding 2 mm. are occasionally present, especially in Division A1 between Beer Head and Salcombe Mouth and in Division C in the Hooken Landslip. A subordinate number of striated prismatic grains with rounded or jagged terminations are always present and range up to 2 mm. in length. Divergent aggregates of acicular tourmaline, closely resembling those occurring in luxulyanite, have been observed in a few samples.

Nearly all the tourmaline is yellow or brown in colour and strongly pleochroic: e, pale yellow; o, various shades of reddish-brown, tawny or orange-brown and yellowish brown. Parti-coloured grains of the dominant yellow/brown variety with patches displaying the pleochroic scheme: e, pale blue; o, rich blue are always present and every gradation exists between these and scarce, completely blue, grains. The blue patches in the parti-coloured grains are usually irregular in shape but the following examples of regular zoning have been observed: (a) prismatic grains of light yellow/orange-brown tourmaline with transverse bands of blue tourmaline arranged at intervals along the length of the grain, *normal* to the c-axis; (b) similar prismatic grains with stripes of blue tourmaline *parallel* to the c-axis; (c) triangular basal sections with a blue rim surrounding an orange-brown core. The frequency of parti-coloured and blue grains is remarkably constant throughout the Cenomanian deposits of south-east Devon (Table II A).

Another distinctive variety of tourmaline has been observed in every sample but rarely exceeds 0.5% of the total tourmaline. This variety possesses the pleochroic scheme: e, lavender, pale pinkish-violet or brownish-violet; o, very deep indigo-blue to opaque as the result of very strong absorption. The optical orientation of such grains serves to distinguish them from dumortierite, a mineral which the present author has searched for but has not yet encountered in the Cenomanian. A few grains displaying very weak absorption of the o-rays and a pleochroic scheme—e, colourless; o, pale brown or blue—are also present in most samples.

Small inclusions are of frequent occurrence in all the varieties of tourmaline, the commonest being acicular or capillary ? rutile, minute crystals of zircon and black opaque dust (? iron oxide) while small voids are sometimes present.

(*ii*) *Topaz* is present in all the heavy fractions examined (Table II A). It is most abundant in the coarser-grained deposits such as Division A1 of the Cenomanian Limestone between Beer Head and Salcombe Mouth and tends to decrease in frequency eastwards towards Humble Point. Nearly all the topaz grains are sub-rounded, roughly equidimensional irregular fragments with diameters ranging up to 0.2 mm. although somewhat larger grains, up to 0.5 mm. in diameter, occur in the coarser-grained deposits west of Branscombe Mouth. Sub-rectangular basal sections bounded by cleavage-planes are not common and occur most frequently in the finer-grained deposits. Most of the grains have a very rough appearance which is only partly due to their high refractive index. Careful focusing shows that the grains frequently display thin closely spaced lamellae normal to their surfaces, a feature which is probably due to corrosion or etching along the basal pinacoid. Small black inclusions are frequently present in the grains but are never abundant.

(*iii*) *Andalusite* is present in all the heavy fractions examined (Table II A). The maximum diameter of the grains is usually between 0.1 and 0.3 mm. but a few grains with diameters up to 0.5 mm. occur in the coarser-grained deposits. As in the case of topaz, andalusite is most abundant in the coarser-grained deposits and tends to decrease in frequency eastwards.

Andalusite never shows the lamellar etching which characterises the surface of many of the topaz grains. Although they show good relief the grains are more pellucid or glass-like in appearance than topaz despite the similar refractive indices of the two minerals. A further distinctive feature is the wide variety of shapes exhibited by the grains of andalusite.

These may be: (a) sub-rectangular prismatic grains with chipped surfaces almost always length-fast and elongated along the c-axis (Fig. 4, 3–4); (b) grains with square outlines and distinct prismatic cleavage which may be accentuated by corrosion. Such grains yield good negative interference figures and are evidently basal sections (Fig. 4, 1); (c) irregularly shaped grains with comparatively smooth outlines. Unlike topaz these grains are never equidimensional nor well-rounded; (d) grains with jagged, zigzag or toothed margins due to regular corrosion, i.e. corrosion along certain crystallographic planes. Early stages in corrosion are seen in prismatic grains with small tooth-like angular projections or scales aligned parallel to the c-axis. More deeply corroded grains have very jagged outlines with deep angular re-entrants (Fig. 4, 2); (e) grains with deeply embayed irregular outlines. Such grains are frequently carious and can sometimes only be described as consisting of an irregular mesh-work. These grains would appear to represent a more advanced stage in corrosion than that described under (d), the corrosion having eaten into the spaces between the original planes of attack (Fig. 4, 5, 6); (f) small wisp-like grains which presumably represent a very advanced stage in corrosion.

The intricate outlines of the deeply corroded grains suggests that they were not transported in their present condition but were corroded after the deposition of the Cenomanian deposits in which they occur. This conclusion is supported by the fact that the most deeply corroded examples occur in the partially decalcified developments of the Cenomanian Limestone such as Division A1 in the large slipped block in the Hooken Landslip known as Mitchell's Rock. Furthermore, andalusite is less frequent at this locality than in adjacent less weathered exposures of the Cenomanian Limestone, Table II A. Andalusite is often regarded as an unstable mineral in sediments and it is probable that its persistence in the Cenomanian Limestone is due to the original detrital grains having been sealed in a hard, impermeable, calcareous matrix during diagenesis and thereby protected at first from corrosive agencies. It was only after the calcareous matrix had been rendered permeable by weathering and partial decalcification that slow intra-stratal corrosion took place. The nature of the corrosive agency is at present unknown but alkaline groundwaters may have been responsible.

Many of the andalusite grains are pleochroic, the pleochroic scheme being X pale pink, deep pink, rose-red, blood-red or rarely brownish-red, Y and Z colourless. The distribution of colour in these pleochroic grains is usually regular but parti-coloured grains containing irregular patches with more intense pleochroism occur in some samples. Regularly zoned grains are very rare but one pale pink grain from the Cenomanian Limestone at Mitchell's Rock displays a narrow rose-red band normal to the c-axis.

Small black inclusions are of frequent occurrence in the andalusite,

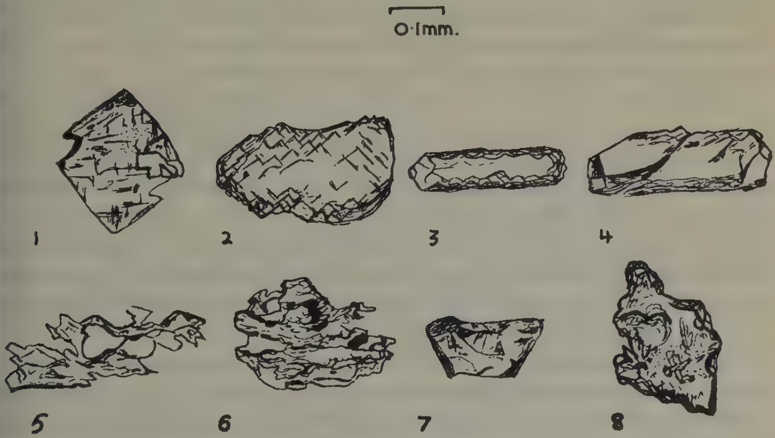


Fig. 4. Andalusite grains from the Cenomanian and Upper Albian deposits of south-east Devon

1. Non-pleochroic basal section. Cenomanian sands of Wilmington facies, Bovey Lane sand-pit, Beer.
2. Grain displaying 'saw-tooth' etching. Cenomanian Limestone A2, Maynard's Cliff.
3. Prismatic grain with chipped margins. Cenomanian Limestone B, Maynard's Cliff.
4. Prismatic grain. Cenomanian sands, White Hart sand-pit, Wilmington.
5. Deeply corroded carious grain. Cenomanian Limestone A1, Branscombe Ebb.
6. Ragged grain. Cenomanian Limestone A1, Branscombe Ebb.
7. Subhedral fragment. Cenomanian Limestone B, Maynard's Cliff.
8. Anhedral fragment. Top of Upper Greensand, Mitchell's Rock (Hooken Landslip).

especially in the non-pleochroic grains, but can rarely be described as abundant. No andalusite with cruciform zoned inclusions, i.e. chiastolite *sensu stricto*, has been observed.

(*iv*) *Zircon.* The zircon content of the heavy fractions varies between wide limits (Table II A). In general, although somewhat sporadic in distribution, zircon tends to increase in frequency eastwards towards Humble Point in contrast with topaz and andalusite. The majority of the grains are either prismatic crystals with fractured or sub-rounded terminations or slightly rounded irregular fragments. Idiomorphic grains with sharp pyramidal terminations are much less frequent and well-rounded ovoid grains are very rare. The long axis of the grains ranges from 0.02 to 0.3 mm. but that of the great majority lies between 0.05 and 0.1 mm. Most of the grains are colourless or greyish. Pinkish and pale purple grains occur sporadically, especially in samples from east of Seaton. The larger grains are usually water-clear and contain small irregularly shaped voids and minute, euhedral, zircon-shaped cavities. Many of the small hypidiomorphic grains contain minute black particles arranged in various ways. These may be uniformly distributed through the crystal as a dust-like suspension or concentrated in the middle as a turbid core or else arranged in narrow zones parallel to the boundaries of the crystal. Many of these dusky zoned zircons resemble those from the Dartmoor granite figured by Brammall & Harwood (1923), Brammall (1928) and Groves (1931).

(*v*) *Staurolite* has been observed in every sample examined but its frequency fluctuates considerably (Table II A). The grains are usually poorly rounded fragments or stumpy, rectangular prisms with more or less rounded edges. They range up to 0.3 mm. in diameter. Weak pleochroism is almost always exhibited, pale yellow to a distinctive golden yellow, often with a slight pinkish tinge. The latter colour is quite unlike any seen among the wide range of yellowish tints exhibited by the tourmaline. Small black opaque ? carbonaceous inclusions are frequently present but can rarely be described as abundant.

(*vi*) *Rutile* has been observed in every sample but tends to be most abundant in the finer-grained residues (Table II A). The grains are either well-rounded and ovoid in shape with diameters up to 0.2 mm. or smaller, rounded, stumpy prisms and fragments. The majority of the grains are reddish-brown, transparent to nearly opaque with a distinctive adamantine lustre, but orange-brown, yellowish-brown and yellowish varieties also occur. Hill (*in* Jukes-Browne, 1903, 274–6) describes two samples of Cenomanian Limestone from the Beer district as containing 'abundant red-brown and orange-yellow fragments of rutile'. At the same time he makes no reference to the most distinctive feature of the heavy residues, viz. the flood of yellow/brown tourmaline. He could not have overlooked these grains and one is forced to conclude that the tourmaline was misidentified as rutile. Rutile is in fact only a minor constituent of all the heavy residues.

(*vii*) *Garnet* is very scarce in the coarser-grained deposits but increases in frequency eastwards. Even so, the heavy fractions of samples taken from the eastern end of the section near Lyme Regis contain only 0.1 to 0.5% of garnet among the transparent grains. It would seem that the frequency of zircon, rutile and garnet in each division of the Cenomanian Limestone is largely controlled by the grade of the sandy constituents as a whole, these minerals tending to be more abundant in the finer-grained residues. The garnet grains are jagged fragments and angular chips up to 0.15 mm. in diameter and colourless or very pale pink. A few grains have been observed with small tooth-like projections and scaly surfaces, features which are characteristic of detrital almandine and usually considered to be an authigenic feature produced by etching.

(*viii*) *Monazite* is of very sporadic occurrence. It is very rare in the deposits west of Beer Head and elsewhere does not normally exceed 0.5% of the transparent grains. Exceptionally high percentages were, however, observed in Division A2 from Pound's Pool Beach, Whitecliff and Humble Point, viz. 1.9, 1.3 and 1.0 respectively.

The monazite occurs as pale yellow zircon-like grains which are occasionally ovoid but more frequently rounded stumpy prisms about 0.05 to 0.15 mm. in length. The identification of the grains as monazite was confirmed spectroscopically.

The following minerals have not been observed in every sample and are always very scarce, their frequency never exceeding 0.5%.

(*ix*) *Cassiterite*. Opaque to sub-translucent grains about 0.1 mm. in diameter and brown, yellowish-brown and reddish-brown in colour. Some of the grains have a distinctive mottled appearance due to irregular patches differing from the groundmass in colour or diaphaneity.

(*x*) *Fluorite*. Angular colourless chips about 0.1 mm. in length.

(*xi*) *Anatase and Brookite*. Anatase is widely distributed but always much less frequent than rutile. The majority of the grains are abraded square plates with sides about 0.1 mm. in length and dark blue or occasionally yellowish-brown in colour. Very rare spindle-shaped opaque greyish grains with an adamantine lustre may be well-worn crystals of octahedrite. Only a few isolated angular fragments of brownish brookite have been observed. With the exception of leucoxene no authigenic titanium minerals have been observed.

(*xii*) *Kyanite*. A single prismatic colourless grain about 0.2 mm. long was found in each of two samples, viz. Division A2 at Whitecliff and Beer Beach. Kyanite must be regarded as being very rare and exceptional in the Cenomanian deposits of south-east Devon.

(*xiii*) *Micas.* Isolated flakes of muscovite and fresh or partly chloritised biotite occur sporadically but are never more than a very minor constituent.

(d) Light Fractions

(*i*) *Quartz.* As a general rule the quartz grains coarser than the coarse sand-grade are well-rounded and often ovoid or spheroidal in shape, the coarse and medium sand-grades are sub-rounded to subangular whilst the finer grades usually consist of slightly rounded angular fragments. Irregularly shaped and jagged grains of the medium and coarse sand-grades are, however, not uncommon in Division A, but it is not clear whether their irregular outlines are always original clastic features. In thin-sections these grains often display deeply embayed margins and are penetrated by the calcareous matrix in a manner suggestive of post-depositional corrosion. No unequivocal proof of such corrosion has been obtained but it is significant that the highly irregular grains appear to be restricted to deposits containing silicified fossils, i.e. those in which diagenetic reactions between calcium carbonate and silica have undoubtedly taken place.

Polycrystalline quartz grains and grains displaying undulose extinction are not common. Polymineralic quartz-aggregates are very scarce although occasional well-rounded grains bearing tar-like excrescences of black tourmaline occur in the coarser-grained deposits. Practically every grain contains minute bubble-like fluid-filled cavities which may be more or less uniformly distributed as a cloud-like suspension, concentrated locally as turbid patches or arranged in lines which are frequently parallel to the c-axis. The commonest mineralogical inclusions are small elongate and stumpy prisms of zircon, opaque particles of red, brown or black iron-ore and irregular clouds of minute greenish chloritic particles.

(*ii*) *Silicified Fossils.* Silicified lamellibranch shells are common in the Cenomanian sands at Wilmington and are often encrusted internally with rings of beekite. Silicified fossils have also been observed in every insoluble sandy residue obtained from Division A but their frequency varies greatly and they rarely possess beekitic encrustations. Divisions B and C rarely contain silicified fossils. The silicified fossils in Division A are mainly fragmentary and indeterminate. Coarsely lamellar shell-fragments referred to the Ostreidae are particularly common and are often associated with costate fragments referred to the Pectinidae. Occasional flat siliceous meshworks, bearing some resemblance to interlocking sponge-spicules, appear to be partially silicified *Inoceramus* shell-fragments. Silicified fragments of smooth, coarsely perforate terebratuloids are fairly common. A few more or less complete valves, one bearing a silicified loop-like brachidium, have been observed and referred to the genus *Kingena*. Fragments of costate brachiopods also occur and appear to be mainly derived from costate terebratuloids such as *Terebrirostra* rather than from rhynchonellids. Small fragments of branching cyclostomatous bryozoa are often present in the insoluble sandy residues. No example of a silicified echinoid has been observed despite the fact that complete tests, isolated plates and spines are common constituents of Division A. Furthermore, none of the ammonites found in this bed show any evidence of silicification.

(*iii*) *Feldspar*. The light fractions contain up to 5% of feldspar fragments which range up to 2 mm. in diameter, the largest fragments occurring in the Cenomanian sands, particularly those in the Bovey Lane sand-pit, Beer. The fragments are rarely well rounded and are mainly sub-rectangular white or pinkish cleavage-fragments possessing a distinctive pearly lustre. In the Cenomanian sands some of the feldspars have a silvery or golden schiller which appears to be due to micaceous inclusions, those possessing a golden schiller are usually pinkish in colour and closely resemble the variety of orthoclase termed murchisonite. No Carlsbad or lamellar twinning has been observed and it would appear that the feldspar grains are predominantly orthoclase fragments derived from comparatively large crystals such as those occurring as phenocrysts in the Armorican granites and quartz-porphyries.

(*iv*) *Glauconite*. Glauconite has been observed in every insoluble sandy residue obtained from the Cenomanian Limestone. In Division A glauconite is rarely apparent in the hand-specimen and the percentage of glauconite is always very small. In Division B and Division C the quartz-glauconite ratio varies considerably even over very short distances but in every case the glauconite must be classed as an abundant and conspicuous constituent. Glauconite is also abundant in the basal Lower Chalk of the Membury outlier but absent from, or extremely scarce, in the Cenomanian sands at Wilmington.

Most of the glauconite falls within the medium and fine sand-grades but slightly larger grains up to 1 mm. in diameter occur infrequently in the coarser-grained deposits, especially Division A1 at Maynard's Cliff. The grains are typically ovoid granules with smooth or irregularly furrowed surfaces and botryoidal aggregates, many of which appear to be internal casts of multilocular foraminifera. Internal casts of small immature helicoid gastropods have also been observed. The glauconite is usually opaque or sub-translucent and when completely fresh varies from very dark green to grass-green in colour but the particularly large grains in Division A1 are often mottled with green and white patches. Grains displaying partial or complete replacement by limonite are common in deeply weathered exposures. In fresher samples the glauconite often possesses a yellowish or colourless outer zone which is more transparent than the core. This form of alteration may extend throughout a grain and appears to be due to the leaching of iron, the glauconite being gradually replaced by a less highly coloured micaceous mineral. The factors controlling this mode of alteration are not clearly understood but it appears to operate in poorly permeable calcareous mosaics where supplies of oxygen are limited, whereas limonitic pseudomorphs are produced by atmospheric weathering of the surface layers.

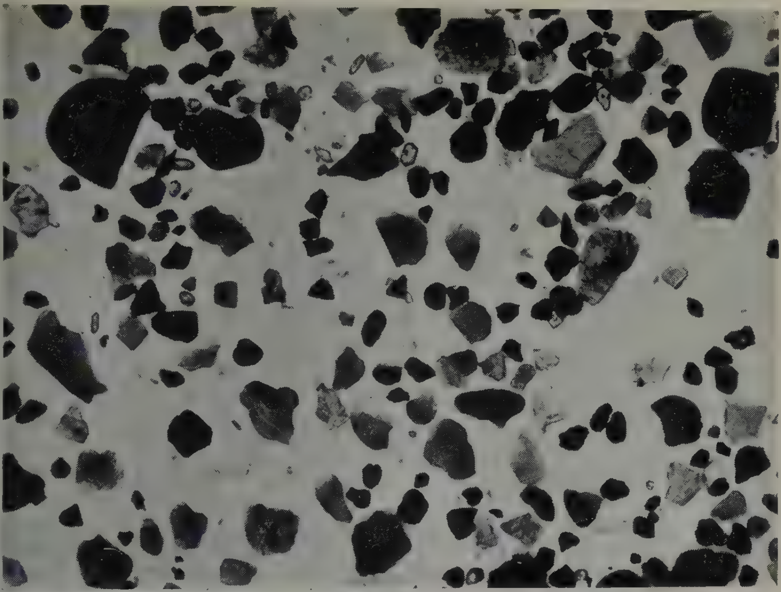
The glauconite is comparatively well-graded and appears to be allochthonous; either transported by currents from an unknown, and perhaps distant, area or passed from one bed to another during periods of localised intra-Cenomanian erosion in the south-east Devon area. Pebbles of limestone bearing a thin greenish, presumably glauconitic, coating occur at the tops of Division A and Division B and such coatings would have been easily removed during transportation and must therefore have developed *in situ*, either by the solution and redeposition of earlier-deposited glauconite or by the direct deposition of primary glauconite. In the present state of our knowledge concerning the mobility of glauconite it is not possible to determine which of these two possibilities is correct.

(*v*) *Serpentine*. A few irregularly shaped grains of yellowish-green serpentine, 0.1–0.2 mm. in diameter, have been observed in all the light fractions from Division A1 but not in the higher beds. The grains could be mistaken for partly altered glauconite but can be distinguished by means of their angularity and coarsely fibrous or mesh-like appearance between crossed-polars.

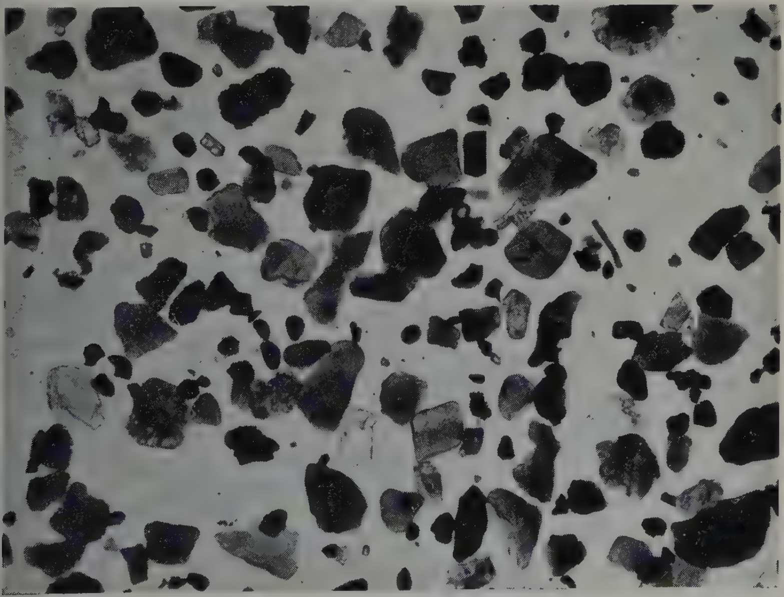
4. COMPARISON BETWEEN THE DETRITAL MINERALS IN THE CENOMANIAN DEPOSITS AND THOSE IN THE UPPER GREENSAND AND MIDDLE CHALK

(a) The Top Sandstones (Table IIB)

The heavy and light fractions from both the Coarse Band and the pebbly sandstone which locally underlies the Cenomanian Limestone have a very close resemblance to those from the Cenomanian deposits (Plate 12B). At least 75% of the transparent heavy minerals in each residue is yellow, orange-brown tourmaline ranging up to spheroidal grains 3 mm. in diameter. Large grains of topaz and pleochroic andalusite are also invariably present, although decreasing in frequency eastwards. An easterly increase in the zircon content is also apparent and appears to be due to the easterly increase in the importance of the finer sand-grades. This conclusion is supported by the fact that the finer-grained, better-sorted sandstones comprising the greater part of the Top Sandstones are consistently rich in zircon, although tourmaline is always the dominant heavy mineral. Muscovite flakes are more frequently present in the Top Sandstones than in Cenomanian deposits, but it is quite clear that the uppermost part of the Upper Greensand is very closely linked with the Cenomanian with regard to the nature and provenance of its detrital minerals.



A



B

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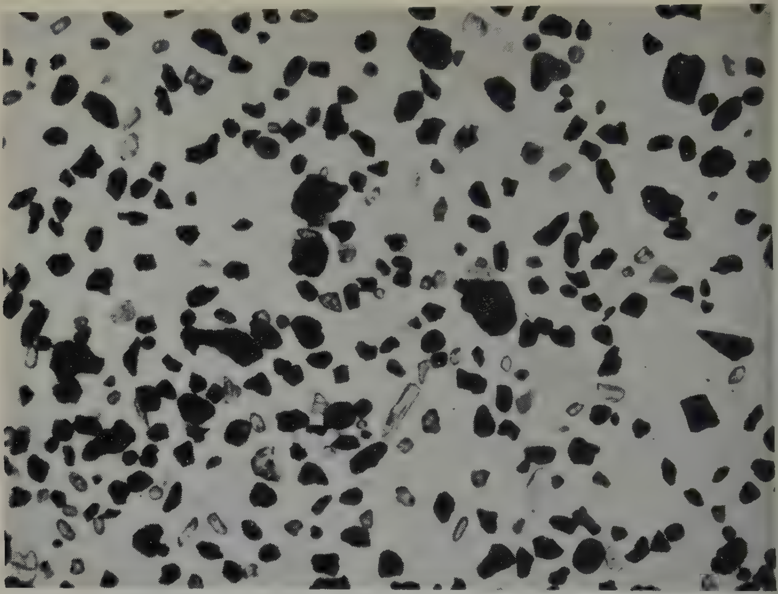


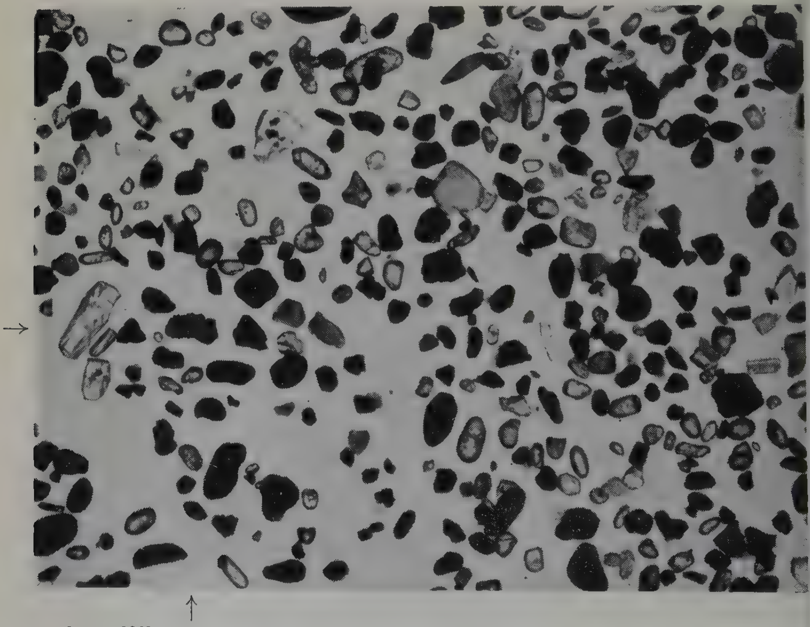


TABLE IIB. **The Principal Transparent Heavy Minerals in certain Upper Greensand Deposits in South-East Devonshire, expressed as Percentages of the Total Transparent Grains, for Comparison with Table IIA**

SAMPLE	*Tt	Tb	Tp	An	Zi	St	Ru	Ky	Si	Ga	N
Top Sandstones											
Berry Cliff	95	3.8	1.9	1.0	—	1.0	1.0	—	—	—	1152
Mitchell's Rock	85	4.0	3.2	0.8	6.5	2.7	1.3	—	—	—	698
Beer Beach (West Side)	79	5.2	1.2	0.4	14.9	3.6	1.0	—	—	—	865
Whitecliff (King's Hole)	76	5.9	1.3	0.4	16.7	4.0	1.1	—	—	—	1253
Humble Point	76	3.4	0.9	0.4	18.6	2.2	2.0	—	—	—	837
Coarse Band											
Berry Cliff	89	7.6	2.1	1.5	3.0	2.7	1.1	—	—	—	372
Beer Head	86	2.8	3.3	—	12.0	0.8	1.3	—	—	—	343
Pound's Pool Beach (south of Beer)	75	3.9	1.2	0.6	20.3	0.6	2.1	—	—	—	592
Chert Beds											
Whitecliff (King's Hole)	69	2.7	⩽0.1	⩽0.1	27.6	1.2	1.6	—	—	—	1629
Corbin Rocks	50	2.6	⩽0.1	⩽0.1	46.2	0.6	3.7	—	—	—	1529
Combe St. Nicholas (near Chard)	54	9.7	⩽0.1	⩽0.1	41.4	1.8	1.8	—	—	—	1290
Coarse glauconitic seam at or near base of Chert Beds											
Little Beach (west of Beer Head)	9.1	1.3	*ABS*	*ABS*	76.1	3.5	7.8	1.8	0.4	1.3	1926
Corbin Rocks	14.1	1.0	*ABS*	*ABS*	68.0	5.5	8.3	3.1	—	1.0	1664
Foxmould											
Peak Hill near Sidmouth	9.5	0.8	*ABS*	*ABS*	88.7	0.2	1.7	0.2	—	—	2550
Whitecliff	11.7	1.2	*ABS*	*ABS*	75.0	3.5	7.0	2.8	—	—	1644
Corbin Rocks	14.6	1.2	*ABS*	*ABS*	64.1	3.2	8.5	4.4	0.6	4.7	1916
Black Ven, near Lyme Regis	6.4	0.8	*ABS*	*ABS*	86.3	1.1	3.3	2.2	0.6	—	3174

* Explanation of symbols: Tt: Total tourmaline, Tb: Blue and parti-coloured tourmaline, Tp: Topaz, An: Andalusite, Zi: Zircon, St: Staurolite, Ru: Rutile, Ky: Kyanite, Si: Sillimanite, Ga: Garnet, N: Total number of grains counted.

(b) The Chert Beds (Table IIB)

The meagre insoluble sandy residues from the Chert Beds consist mainly of angular or subangular fragments of quartz, rarely exceeding 0.25 mm. in diameter. Glauconite granules and muscovite flakes are also invariably present but vary considerably in abundance from bed to bed.

In order to obtain heavy fractions comparable in bulk to those obtained from the higher formations it is necessary to treat very large samples, weighing at least five hundred grams. The heavy fractions obtained in this way contain more zircon and ferruginous opaque grains than the Top Sandstones. The frequency of tourmaline decreases eastwards in sympathy with an increasing zircon-content but the yellow/orange-brown variety of tourmaline is always very abundant. Furthermore, the majority of the heavy minerals in the Chert Beds are in very close varietal agreement with those found in the higher Cretaceous rocks. The sand in the Chert Beds is evidently fine-grained detritus derived from a similar source to that providing the sand in the Top Sandstones and Cenomanian deposits.

(c) The Foxmould

A preliminary study of the Foxmould (Table IIB) shows that it contains a heavy mineral suite distinct from those in the overlying Cretaceous rocks

(Plate 13A). The most significant differences are (a) the dominance of zircon, including many well-rounded polycyclic grains, (b) the yellow/orange-brown variety of tourmaline rarely exceeds 15% of the transparent grains, (c) minerals such as kyanite, sillimanite and garnet, which are either very scarce or completely absent from the higher beds, are often not uncommon, (d) the presence of mineral varieties not so far recorded above the Foxmould. Noteworthy among these grains is a pale yellow, mottled variety of tabular anatase which sometimes occurs as polycrystalline aggregates. It may be significant that similar grains have been recorded in the New Red Sandstone of south Devon (Thomas, 1902, 1909).

Coarse-grained seams of glauconitic sand at or near the junction with the Chert Beds contain heavy minerals similar to those in the underlying Foxmould (Plate 13B). The most distinctive constituents are: abundant large colourless or purplish ovoid grains of zircon up to 0.20 mm. in diameter; large rectangular and square, cross-fractured prisms of kyanite up to 0.4 mm. in length; colourless and pinkish fragments of garnet and brownish staurolite, distinct from that occurring in the higher beds and often displaying saw-tooth boundaries. Topaz and andalusite appear to be completely absent whilst tourmaline comprises only about 10% of the transparent grains. The presence of the Foxmould suite in these *coarse* greensands shows that the contrast between the heavy minerals in the Foxmould and the overlying Cretaceous rocks is not due to grading. Taken in conjunction with the big differences between the overall lithology of the Chert Beds and that of the Foxmould the contrast must reflect a rapid and fundamental palaeogeographical change.

(d) The *Inoceramus labiatus* Zone

The low sand-content of the *I. labiatus* Zone prevents a reliable determination of the relative frequencies of the detrital minerals contained in it. All the minerals occurring in the Cenomanian deposits have, however, been found in the *I. labiatus* Zone, glauconite, quartz and tourmaline being invariably present. All the mineral grains are in complete varietal agreement with those in the Cenomanian deposits and are most frequent towards the base of the zone and close to areas, such as Branscombe Mouth, in which Cenomanian deposits were undergoing erosion during early Turonian times. It is therefore not necessary to postulate an influx of sand from outside south-east Devon; all the grain in the Middle Chalk of this area could have been introduced by the local reworking of Cenomanian deposits. The chromite and serpentine in the Beer Stone (Macdonald Davies, 1919) could have been derived from Cenomanian deposits in the Branscombe Mouth region; they do not prove that the Lizard serpentinites, or similar rocks, were being eroded during *Turonian* times.

5. THE INTERPRETATION OF THE DETRITAL MINERAL ASSEMBLAGES IN THE UPPER ALBIAN, CENOMANIAN AND EARLY TURONIAN DEPOSITS

(a) The Foxmould

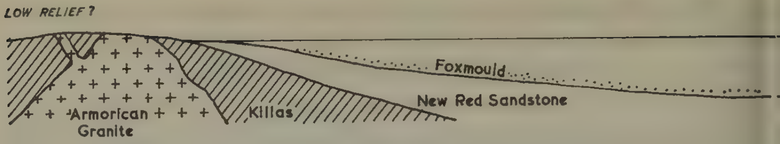
The detrital minerals in the Foxmould appear to contain a high proportion of grains derived from pre-Cretaceous sediments. Much of this detrital material, including minerals of undoubted Armorican type, could indeed have been derived from the New Red Sandstone across which the Foxmould is strongly transgressive (Fig. 5, I). The kyanite, however, displays no evidence of the formation of shimmer-aggregates (Thomas, 1902) and it would appear that the Foxmould contains some first cycle detritus including sillimanite and garnet in addition to kyanite. The location of the area of metamorphic rocks from which these grains were derived has so far not been determined but the small amount of evidence at present available suggests that they may have been of easterly derivation. Any *direct* contribution made by the Armorican granite-complexes of Devon and Cornwall appears to have been very small.

(b) The Chert Beds

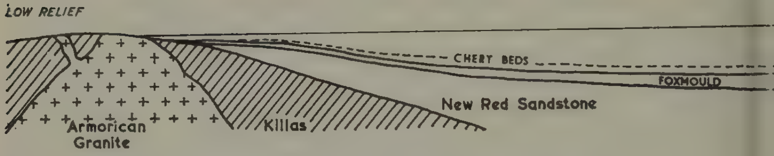
Calcite is not common in the Foxmould apart from that in the concretionary layers or ‘burrs’. On the other hand calcite is the dominant constituent of the Chert Beds, their sand-content being very small. This sudden change in lithology is most readily explained by postulating that the rocks supplying the detritus to the Foxmould were submerged beneath the sea by an Upper Albian transgression (Fig. 5, II). The Upper Albian sea evidently made extensive contacts with a landmass or river-borne sediments which were later to provide an increased supply of detritus to the very late Albian and early Cenomanian deposits. A discussion of the nature and location of this landmass will be given after a consideration of the detrital minerals of these later deposits. The small sand-content of the Chert Beds, however, appears to indicate that the landmass was low lying and not undergoing vigorous erosion during the accumulation of these beds.

(c) The Top Sandstones, Cenomanian Sands of Wilmington Facies and Division A1 of the Cenomanian Limestone

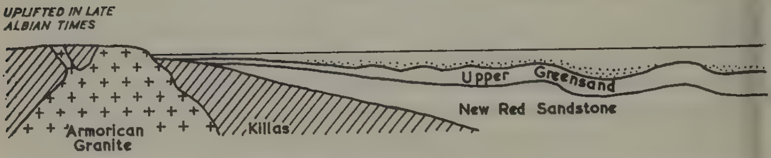
The nature of the inclusions and the scarcity of strain-shadows in the quartz, the occurrence of orthoclase, including the variety murchisonite, and the varieties of zircon, rutile, anatase and monazite suggest that granitic rocks provided the bulk of sand-grains in the Top Sandstones and early Cenomanian deposits. The dominant heavy minerals, namely the variously coloured varieties of tourmaline, the topaz and the scarce



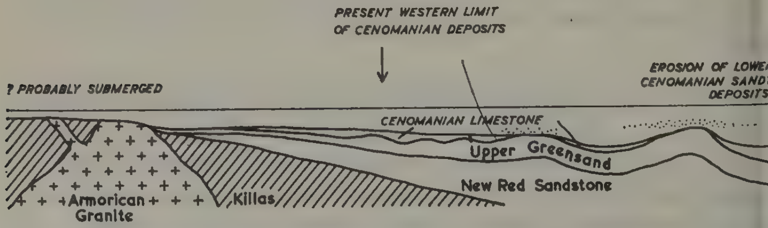
I DURING ACCUMULATION OF FOXMOULD



II DURING ACCUMULATION OF CHERT BEDS



III DURING ACCUMULATION OF EARLY CENOMANIAN SANDS OF WILMINGTON FACIES AND DIVISION AI



IV DURING ACCUMULATION OF INOCERAMUS LABIATUS ZONE OF THE MIDDLE CHALK fluorite and cassiterite are all in close varietal agreement with the constituents of pneumatolytic rocks and mineral-veins associated with the Armorican granites of Devon and Cornwall. Furthermore, clear strongly pleochroic andalusite is especially characteristic of some of the Cornish granites, notably Bodmin, St. Austell and Falmouth (Carnmenellis) (Ghosh, 1927, 1928).

The lateral variations in sand-content and grading confirm that the sand-grains were derived from the area in which Armorican granites are emplaced, i.e. an area lying to the west or south-west of south-east Devon.

The only minerals which could not have been derived from the Armorican granite-complexes are staurolite and serpentine. The staurolite never displays saw-tooth etching; this fact and the relatively low degree of rounding displayed by the staurolite suggests that it was derived directly from metamorphic rocks and not from pre-Cretaceous sediments such as the New Red Sandstone (Thomas, 1902). Tilley (1937) records a similar variety of staurolite as a constituent of certain cordierite–anthophyllite granulites in the Old Lizard Head Series at Pistil Ogo. The present outcrop of staurolite-bearing rocks in the Lizard complex is very small but a former extension, now covered by the English Channel, would appear to be the nearest possible source of the staurolite in the late Albian and early Cenomanian deposits. A minor contribution by rocks of the Lizard complex is certainly suggested by the presence of serpentine fragments in Division A1.

Although grains derived from pre-Cretaceous sedimentary rocks are very scarce in the Cenomanian deposits this does not mean that no such rocks were exposed to denudation during late Albian and early Cenomanian times. The New Red Sandstone probably made no significant contribution because it was almost, if not entirely, covered by earlier Cretaceous sediments. On the other hand the Devonian limestones, slates and grits and the Culm could have been widely exposed without making any important contribution of coarse-grained clastic material, any contribution being swamped by that provided by much smaller granitic exposures.

Fig. 5. Diagrammatic sections showing principal palaeogeographical changes during the accumulation of the Albian, Cenomanian and early Turonian deposits of south-east Devon

I. Transgressive Upper Albian sea. Detrital minerals derived mainly from the New Red Sandstone.
II. New Red Sandstone largely removed by erosion or buried beneath marine deposits, scarce detrital minerals derived mainly from Armorican granite-complexes.
III. Early Cenomanian deposits laid down on a current-scoured surface of Upper Greensand, some irregularities on sea-floor probably due to small-scale tectonic activity. Detrital minerals mainly derived from Armorican granite-complexes.
IV. Transgressive Turonian sea. No direct influx of detritus from Armorican granite-complexes, sand-grains derived from early Cenomanian and late Albian deposits undergoing erosion in areas of small-scale anticlinal uplift.

The closest granite-mass to south-east Devon is that of Dartmoor and there would appear to be little doubt that this complex made important contributions of detrital grains to the late Albian and early Cenomanian seas covering south-east Devon. The dusky and zoned zircons and certain other heavy minerals in these deposits closely resemble those from the Dartmoor granite and detritals figured by Brammall & Harwood (1923a, 1923b), Brammall (1928) and Groves (1931). The writer agrees with Wells (1931), however, in stressing that the varieties of zircon and other heavy minerals gives no precise indication of derivation from a particular granite-mass; the varietal characteristics of such minerals may well vary with depth. All the Armorican granites must be more deeply eroded than in Cretaceous times, indeed some may not have been exposed at all until after Cretaceous times. In the present state of knowledge, therefore, it is precipitate to refer any of the detritals in the Cenomanian to one particular complex. For example, a contribution by one or more of the Cornish granites is not conclusively proved by the common occurrence of clear strongly pleochroic andalusite such as occurs in the present-day exposures of these granites, despite its relative scarcity in the Dartmoor granite. The scarcity of chiastolite suggests that contaminated granites rather than aureole-rocks were the main source of the andalusite. At the present day andalusite is restricted to the marginal facies of the Dartmoor granite (Brammall & Harwood, 1923b). Nevertheless, since the level reached in the erosion of the Dartmoor mass is probably lower than in the Cornish granites, andalusite-bearing contaminated granites emplaced nearer the original roof of the Dartmoor intrusion may well have been widely exposed in late Albian and early Cenomanian times. Subsequent erosion has completely removed this contaminated zone, apart from a narrow marginal belt in contact with the country-rock.

An assessment of the relative importance of any contribution by the Cornish granites is also rendered almost impossible by the fact that the bulk of the detrital grains, whilst typically Armorican, could have been derived from any of the granite-complexes. A contribution by the Cornish granites is, however, suggested by the wide variation in the degree of rounding of equivalent grade-sizes of the quartz and tourmaline. This variation is probably due to varying distances of transport so that it is reasonable to suppose that contributions were made by granite-masses lying farther afield than Dartmoor, not only from the Cornish granites but also from others which may be submerged beneath the English Channel or Western Approaches. The various contributions could have been mingled during transportation into south-east Devon by a process such as along-shore drifting in a north-easterly or easterly direction.

The increased influx of sand which led to the accumulation of the Top Sandstones above the highly calcareous Chert Beds was probably due to uplift of the rocks comprising the Cornubian massif. These earth movements could ultimately have led to the development of very shallow-water conditions in south-east Devon leading to the widespread erosion which preceded the accumulation of the Cenomanian deposits throughout that region. The second influx of sand, i.e. that found in Division A1 and the sands of Wilmington facies, would then be due to a Cenomanian transgression towards the Cornubian massif, accompanied by sand-deposition on the margins of a shallow sea (Fig. 5, III).

A somewhat surprising feature of the clastic grains in the late Albian and early Cenomanian deposits is that, despite the abundance of granitic debris, biotite and its undoubted alteration products are very scarce. Biotite is abundant in modern sediments adjacent to the Armorican granites. For example, the present author has found that the modern shell-sands flanking the Land's End and Scilly granites have sand-contents and insoluble sandy residues resembling those from the Cenomanian deposits, except that biotite, instead of tourmaline, is the dominant heavy mineral. It is not impossible, therefore, that biotite, and perhaps other unstable ferromagnesian minerals, were once present in the late Albian and early Cenomanian deposits of south-east Devon but were decomposed during diagenesis. The early decomposition of these ferromagnesian minerals provides an explanation for the presence of limonite sealed in an apparently unweathered calcareous mosaic and also provides a possible source for the silica responsible for the silicification of calcareous organic remains in the early Cenomanian deposits. A full discussion of the source of this silica will be given in a subsequent paper but it is noteworthy that the above explanation accounts for the scarcity of silicified fossils in the higher beds. Their scarcity could be due to the absence of any further significant influxes of fresh granitic debris, the unstable ferromagnesian minerals in the reworked deposits having been decomposed before they came under the influence of erosive agencies.

(d) Divisions A2, B and C of the Cenomanian Limestone, the Membury Lower Chalk and the Middle Chalk

Most, if not all, of the detrital grains in the higher divisions of the Cenomanian Limestone appear to have been derived from early Cenomanian or late Albian deposits. It is not possible to prove that there was *no* direct contribution from the Armorican granites during Lower Cenomanian times after the accumulation of Division A1. Nevertheless, the purity of the Lower Chalk above the Basement-bed at Membury and the Middle Chalk when not adjacent to centres of intra-Cretaceous tectonic activity shows that no significant direct contribution was made during Upper Cenomanian and Turonian times. This cutting off of the direct supply of detritus from the Armorican granites is most readily explained as the result of a continuing Cenomanian transgression, the granites being probably completely submerged beneath the sea during Upper Cenomanian and Turonian times (Fig. 5, IV).

The detrital minerals in both the phosphatised pebbles and the matrix of the Chalk Basement-bed at Membury and in adjacent areas in Somerset and Dorset closely resemble those in the Cenomanian Limestone. In these areas absence of sand in all but the lowermost few feet of Lower Chalk and the rich derived fauna in the Basement-bed (Hancock, *in* Smith, 1957b) suggest that the sand and granitic detritus which entered south-east Devon during early Cenomanian times spread beyond the limits of the county. In the Membury district, as in adjacent areas of Somerset and Dorset, these early Cenomanian sandy deposits were subsequently removed by erosion, some of the sand being incorporated in the Basement-bed and lowermost Lower Chalk by a complex sequence of erosion, reworking and deposition extending into Upper Cenomanian times.

ACKNOWLEDGMENTS

The author wishes to thank Dr. W. F. Fleet and Mr. D. J. Shearman for their helpful criticism of the contents of this paper. He also wishes to thank Mrs. G. Cook for the drawings of andalusite, shown in Fig. 2.

EXPLANATION OF PLATES

PLATE 12

A. Heavy minerals, Cenomanian Limestone, Division A2; west side of Beer Beach
B. Heavy minerals: Top Sandstones, immediately below the Cenomanian Limestone. Division A2, west side of Beer Beach.
Both ×27, plane-polarised transmitted light.

Note the flood of tourmaline and the similarity in the grade-sizes and the rounding of the grains in the two suites. A few small euhedral grains of zircon can be seen in Plate 13A

PLATE 13

A. Heavy minerals, average sample of Foxmould; Corbin Rocks, east of Culverhole Point.
×27, plane-polarised transmitted light.

Note the abundance of opaque grains and small grains of zircon, most of the latter being well-rounded. A prismatic grain of kyanite can be seen just below the centre of the photograph

B. Heavy minerals from pebbly seam of greensand, base of Chert Beds, Little Beach, west of Beer Head.
×27, plane-polarised transmitted light.

Note the abundance of well-rounded grains of zircon and the presence of prismatic grains of kyanite, the largest of which is on the left-hand side of the photograph.

A rectangular grain of staurolite with 'saw-tooth' margins can be seen about half an inch to the right of the largest grain of kyanite (The co-ordinates of the staurolite are indicated by arrows)

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DISCUSSION

Dr. J. M. Hancock, in a written contribution, wrote that Dr. Smith is to be congratulated on what must be the most detailed study to date of the heavy minerals of a marine formation in Britain. The writer is himself working on the heavy minerals of the Chalk basement beds throughout southern England, and has obtained similar results in Devon. He agrees with the westerly and local origin of the sand but not with the exposure and erosion of the Dartmoor granite. The very high tourmaline to zircon ratio and the murchisonite variety of orthoclase suggest erosion of the tourmalinised envelope of the granite and its offshoots. The characteristic Dartmoor zircons have already been recorded from the New Red Sandstone by Groves (1931), and the late Dr. Hutchins (in an unpublished thesis) found staurolite in the New Red Sandstone of the Crediton Valley. One staurolite grain from Wilmington seemed to have outgrowths which are themselves slightly worn and rounded, and this grain could hardly be from a primary source. It seems possible that the New Red Sandstone was a more important source than Dr. Smith suggests. Not only would it have been a more extensive formation than now, but it may be questioned if the early Cenomanian actually represents a transgression which would have protected the New Red Sandstone from sub-aerial erosion.

The writer's work in Dorset, Wilts and Hants shows that farther out to sea the detritus was derived from a wider variety of sources.

THE AUTHOR expressed his pleasure at hearing that Dr. Hancock was in general agreement with his conclusions. Grains derived from the Armorican rocks of Cornubia are undoubtedly present in the New Red Sandstone of south-west England. One cannot therefore entirely rule out this formation as a source of some of the early Cenomanian detritals, especially single grains. Nevertheless, the composition of the Cenomanian detrital assemblages contrasts very strongly with those both in the New Red Sandstone and in the Foxmould and it would appear that any contribution made by the New Red Sandstone to Cenomanian sediments was very small when compared with the great flood of granitic detritus. The outlines of many of the staurolite grains are not in agreement with the suggestion that all the staurolite was derived from older sediments.

The author stated that he did not consider that the very high tourmaline to zircon ratio necessarily indicates that the granites proper were not exposed. The zircon content of the Cenomanian sandy detritus increases eastwards, i.e. in the same direction as the detritus becomes finer-grained, and it would appear that the ratio is profoundly influenced by the lateral grading and distance of transport. Furthermore, the present outcrops of the Armorican granites are not all rich in zircon, this mineral being scarce in the Land's End granite and in the adjacent shoresands. The possibility that the zircon content of a granite-mass varies with depth must also be borne in mind.

With regard to the broader questions, apart from biotite, heavy minerals in the Armorican granites proper are far less numerous than in pneumatolytic rocks such as schorl-rock. Consequently far smaller exposures of pneumatolytic rocks would be required to give a major contribution of heavy minerals than in the case of the granites themselves. It is true that biotite might have been expected as a frequent heavy mineral in the early Cenomanian deposits if the granites were widely exposed during their accumulation. In the paper it is suggested that this may indeed have been the case but unlike tourmaline, most of the biotite was destroyed by post-depositional decomposition.

Much depends upon one's definition of the granites proper, as opposed to offshoots and envelopes, but there would appear to be no clearly defined criterion by which to estimate the depth of erosion of the Armorican granite-complexes. The smaller present-day outcrops of these granites were probably not laid bare in Cretaceous times; the larger must have been less deeply denuded. It would, however, be preferable to quote Armorican granite-*complexes*, and not the granites themselves, as the main source of detritus in the early Cenomanian sediments.

The Nature and Origin of Chert in the Upper Greensand of Wessex

by G. R. TRESISE

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ABSTRACT: The sporadic distribution of chert, both vertically and laterally, within the Upper Greensand, is described. The chert beds of Devon and the Dorset coast consist of glauconitic sands, usually calcareous but leached free of calcareous material wherever there is no protective capping of Chalk. In contrast the chert beds of Somerset and the Vale of Wardour consist of sparsely glauconitic silty sediments. The chert nodules themselves typically show a massive flinty core surrounded by a porous light-coloured marginal rock.

The various theories of formation previously forwarded for these cherts are examined, but appear inadequate to account for all their features. It is postulated that decomposing organic material caused the precipitation of masses of silica gel on the Cretaceous sea-bed; that these dehydrated into nodules of hard but porous stone, which were then converted into true chert by redistribution of the silica of opaline sponge spicules. A complementary migration of calcareous material often produced doggers of limestone in the intervening sands. Slight variations of acidity, or possibly of temperature, within the sediment are thought to control the process.

1. INTRODUCTION

PROBABLY NO OTHER TOPIC of sedimentary petrology has been so assiduously studied and so copiously documented as has the problem of chert formation. Probably too in no other case has so exhaustive a study produced so little agreement on the salient facts. Whether cherts are organic or inorganic in origin, contemporaneous or replacive features, deposited *in situ* or redistributed in solution, are issues still in dispute. In the great volume of geological literature which has been produced on the subject, those cherts which occur in beds of limestone have received most attention. However, one of the classic theories of chert formation—that put forward by Hinde in 1885—was based on the cherts present in two arenaceous divisions of the Cretaceous system. It is with the cherts of one of these beds, the Upper Greensand of Albian age, that the present paper is concerned.

Hinde claimed that the cherts of both the Lower and the Upper Greensand originated from dense felted masses of the spicules of siliceous sponges, cemented by chalcedonic silica deposited from solution. Jukes-Browne (1900), in his monumental account of the Upper Greensand, accepted this view without question, and equated the sporadic occurrence of chert beds with the patchy distribution of sponge-fields on the present sea-floor. Later workers on the Upper Greensand either omitted all reference to chert formation or else accepted Hinde's interpretation. In recent years, however, the validity of these views as applied to the Lower Greensand has been questioned first by Richardson (1947) and later by Humphries (1956). The time seems ripe, therefore, for a reconsideration of the features shown by the cherts of the Upper Greensand, and of the conclusions drawn as to their mode of formation.

Siliceous Stone in the Upper Greensand

Chert has recently been defined by Smith (1960, 1) as 'a siliceous rock consisting predominantly of microcrystalline or cryptocrystalline non-clastic silica and occurring as beds or as nodular masses or veins in sedimentary rocks'. This definition excludes the siliceous sandstones where the greater part of the silica is in the form of clastic grains of quartz, and non-clastic silica plays a subordinate role as cement.

Siliceous sandstone is present in the Upper Greensand beds of the Blackdown Hills and the Vale of Wardour, but invariably occurs in sands which have been leached free of calcareous material. Limestone doggers are never present, and even where original calcareous shells appear to be preserved, they have invariably been replaced by silica. It seems very probable that the siliceous sandstone is produced by a similar replacement of calcareous concretions and is therefore of secondary origin, whereas the true chert shows every indication of being a primary feature.

2. DISTRIBUTION OF CHERT IN THE UPPER GREENSAND

The Albian sediments of south-eastern England are of two distinct facies, with dark-coloured Gault Clays at the base passing up into a series of 'greensands'. The latter thicken westward and in Devon and Somerset are the sole representative of the Albian. This so-called 'Upper Greensand' is composed of a series of glauconitic sands which contain nodules of chert, limestone and phosphatic material. Chert is the most sporadically distributed of these components, and the accompanying maps (Figs. 1 and 2) show the approximate limits of its extent. Chert beds are best developed in western Wessex and thin out when traced east. In east Dorset and in the Devizes (41/004605) district of Wiltshire the Upper Greensand succession is completely free from chert, though it reappears in the Isle of Wight. The

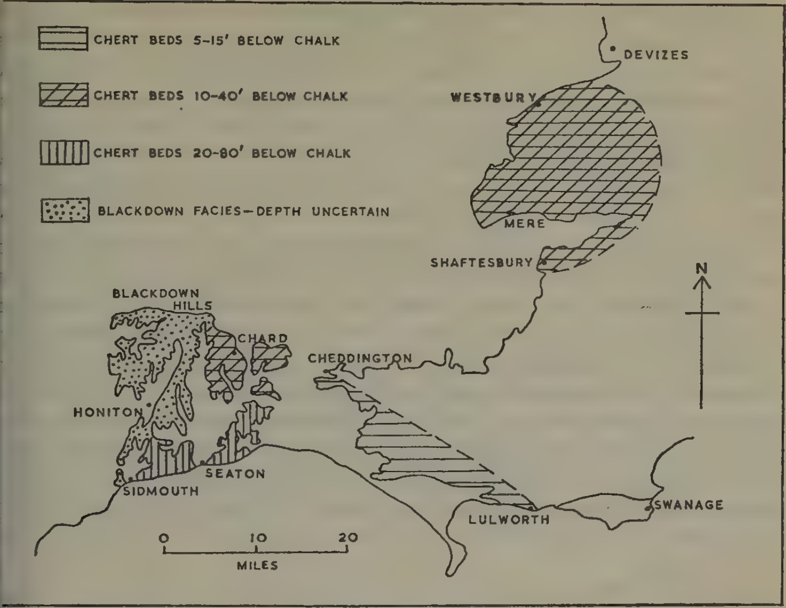

CHERT BEDS 5–15′ BELOW CHALK
CHERT BEDS 10–40′ BELOW CHALK
CHERT BEDS 20–80′ BELOW CHALK
BLACKDOWN FACIES—DEPTH UNCERTAIN
DEVIZES
WESTBURY
MERE
SHAFTESBURY
N
BLACKDOWN HILLS
CHARD
CHEDDINGTON
HONITON
SEATON
SIDMOUTH
LULWORTH
SWANAGE
0
10
20
MILES

Fig. 1. Distribution of Chert in the Upper Greensand related to the base of the Chalk

latter probably represents a separate basin of deposition with sediment derived from Brittany and the Cotentin Peninsular (Groves, 1931) whereas in Wessex the source was a landmass to the west.

In addition to its limited areal extent, chert is also restricted within the vertical sequence of Upper Greensand beds. In general it characterises the higher part of the succession, though it is always absent from the highest few feet of the division. Fig. 3 shows the vertical distribution of chert within the Upper Greensand at selected localities in Wessex, together with the rather meagre evidence provided by Spath (1921–43) as to the place of these beds in his sub-zonal scheme.

On the coast, chert beds attain their maximum development, almost seventy feet, in the *substuderi* Zone at Seaton (30/235896). To the west, evidence of current activity becomes increasingly pronounced in the higher beds, and chert is correspondingly restricted to the lower horizons of the zone. Even at Seaton eighteen feet of chert-free beds immediately underlie the Chalk. As the beds are traced east into Dorset this process is reversed, and chert is progressively confined to the higher horizons. At White Nothe (30/770811) and Durdle Door (30/803803) six feet of chert beds occur in the *dispar* Zone a few feet below the Chalk. Chert finally disappears from the succession just east of Lulworth (30/826801).

Inland chert occurs in fine-grained silty sediments at similar levels below the Chalk in both the Chard (31/313088) district of Somerset and the Vale of Wardour. Again chert is lacking from the topmost beds, this being most pronounced in the west. Chert beds are most extensively developed in the Blackdown Hills district, though the highest Upper Greensand beds, together with the overlying Chalk, have here been stripped off by erosion. Spath (1921–43) showed that these chert beds were of *varicosum* age and so pre-dated those of more easterly areas.

Chert thus occurs at widely varying horizons in different areas (Figs. 1 and 2). Chert formation was first initiated in the Blackdown Hills region, at a time when Gault Clay was still being laid down over much of Wessex. At a later stage it spread to the Devon coast, the Chard district of Somerset and the Vale of Wardour. Finally, it extended to the Dorset coast west of Lulworth. It will subsequently be shown that chert formation was favoured by tranquil waters, and this migration is therefore thought to reflect the increase in current activity which resulted from the shallowing of the seas towards the close of the Albian period. The culmination of this process may account for the absence of chert from the very highest beds of the Upper Greensand.

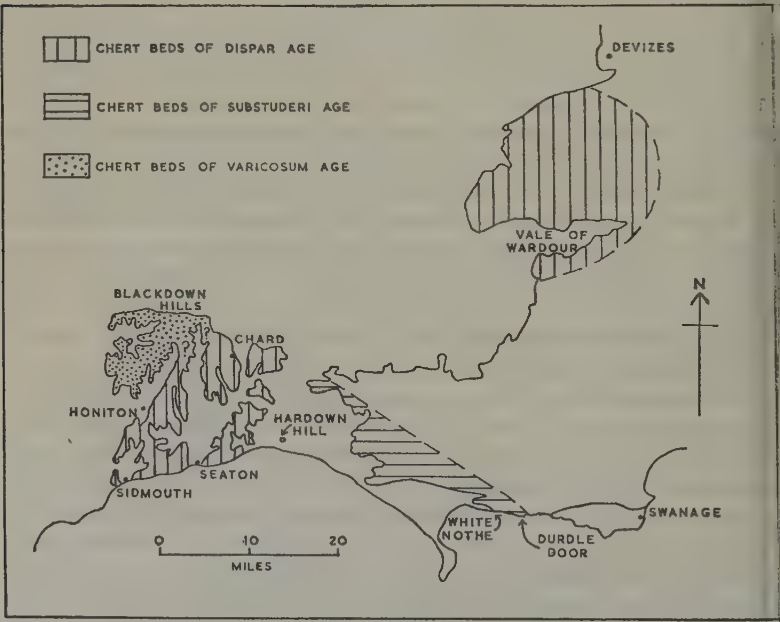


Fig. 2. Distribution of Chert in the Upper Greensand related to the Zonal scheme of Spath

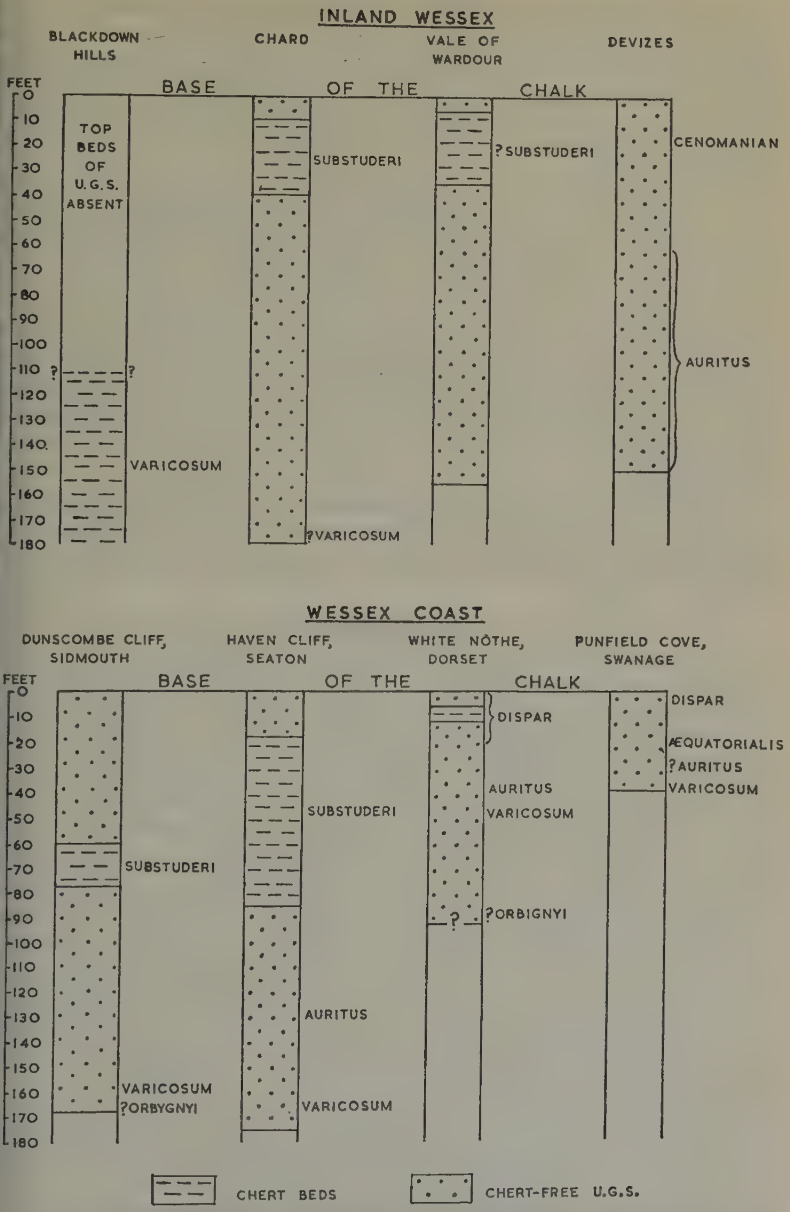


Fig. 3. Horizons of Chert Beds in the Upper Greensand. Base of Chalk as datum

3. LITHOLOGY OF THE CHERT-BEARING BEDS

The Upper Greensand is composed of arenaceous sediments of variable lithology, and this is reflected in the varied nature of the groundmass in which chert nodules occur. Some generalisation is possible, however, both on the type of sediment in which chert is found, and also on the types from which it is absent.

The Chert Beds of the Devon coast consist of buff or brown sands of fine or medium grade, which are glauconitic, ferruginous and usually calcareous. At Hardown Hill (30/402944) and Peak Hill, Sidmouth (30/109867), the sands have been leached completely free from the calcareous material, in the form of fine shell debris, which elsewhere characterises the beds. 'Limonite' tends to be most abundant where the sands are non-calcareous and it thus seems possible that the leaching processes were accompanied by the decomposition of glauconite to iron oxides. The Devon cherts are usually black, with a porous marginal zone stained yellow or brown by limonitic material; iron-staining is also present along fractures in the chert core.

The chert beds of the Dorset coast consist of fine calcareous sands; lack of ferruginous material, rather than an abundance of glauconite, is responsible for their greenish-grey colour. The chert here is usually grey or brown in colour.

The chert beds which occur inland in Wiltshire and Somerset are of quite different lithology. They consist of very fine sands, grading down into silt, which are usually richly calcareous. Often only a few tiny grains of glauconite are present, while limonite may be totally lacking so that the sediments are white and almost chalk-like in appearance, Dead Maid's quarry, Mere (31/803323), providing a good example of this. The cherts characteristic of this lithology are black flint-like types with a pure white marginal rock. The fine grain-size of these sediments seems to imply very weak current activity, since the seas were unlikely to have been of any great depth. In the Blackdown Hills to the west, sands of medium grade containing both glauconite and limonite form the groundmass of the chert beds, the lithology here resembling that of the Devon coast. Leaching has completely eliminated all calcareous material, and any shells still present have been replaced by silica. Siliceous (probably silicified) sandstone, unimportant elsewhere and completely lacking from the coastal sections, is also characteristic of this region.

Chert tends to be absent from glauconitic sands which are almost free from limonite and also sands of orthoquartzitic type deposited in shallow water or during periods of strong current action.

This first group includes the Foxmould sands of Devon and the whole of the Upper Greensand succession at Punfield Cove, Swanage (40/045813), which is of Foxmould lithology throughout. Often, as at Durdle Door

(30/803803), the beds richest in glauconite immediately underlie the chert-bearing horizons where there is a marked decline in the glauconite content. This antipathy between chert and glauconite suggests that conditions favourable to one inhibited the formation of the other. In an earlier paper (Tresise, 1960) the author has argued that, while still waters were essential for chert formation, glauconite seems to have been most abundantly formed during or immediately after periods of marked current activity. A supplementary factor may be the fact that both chert and glauconite require a source of soluble silica so that, if the supply of silica is limited, formation of one would automatically tend to restrict the formation of the other.

The second group includes the highest beds of the Upper Greensand. At Haven Cliff, Seaton (30/257897), the top eighteen feet of the succession are chert-free. These beds tend to approach the orthoquartzitic, both limonite and glauconite being very sparse, although calcareous material is still included. In the underlying beds, pebble bands, current-bedding, shell beds and a penecontemporaneous breccia all indicate strong current activity, with chert confined to the intervening unaffected sands. The dying out of chert in the higher beds, as well as the progressive westward restriction of chert-bearing beds, probably reflect increasing current activity in shallowing waters.

4. FEATURES OF THE CHERT NODULES

Chert in the Upper Greensand occurs either as nodules, or as lenticular bands which run parallel to the bedding for a few feet before dying out. Where the chert beds are best developed, in the Haven Cliff section near Seaton, individual bands are six to nine inches in height and are separated by a foot or more of intervening sand. The bands rarely exceed a foot in height, though a three-foot band of grey chert, admittedly of very limited lateral extent, was present in a fallen block below the cliff. In the higher beds chert is more sporadically developed, and occurs as small irregular nodules or as 'stringers' within the sand. Inland the thickest bands seen were those of the Snowdon Hill quarry, Chard (31/313088), which averaged nine inches to a foot, though again the highest beds were thinner and seldom exceeded six inches.

The individual nodules of chert are not of uniform nature throughout. The core consists of compact glassy chert which may be black, brown, grey or almost white in colour. This is surrounded by a band of porous siliceous rock which may be white but is often stained yellow or brown by iron; this marginal zone is usually a half-inch to an inch in width. Nodules of a similar porous white siliceous stone, lacking the chert core, sometimes occur in the chert beds, notably in the Vale of Wardour.

The features shown by the porous marginal zones of the chert nodules have received little attention in the past, though they throw valuable light on the probable mode of formation. The general characteristics of these rocks may be summarised as follows:

The marginal rock is not transitional between the chert core and the surrounding sediment.

Though the junction between core and marginal rock appears sharp and clear-cut, the two do represent a single unit and cannot be separated in the way that the nodule as a whole may be broken out of the surrounding sediment.

Both chert and marginal rock are dominantly composed of chalcedonic silica.

In proportion and nature, the detrital grains present resemble those included in the core.

The chief distinction from the core is the very porous nature of the rock. Moulded in the chalcedony of the groundmass are cavities of three distinct types:

(a) Original cavities between the botryoidal ('colloform') masses of chalcedony which compose the groundmass;

(b) Solution cavities of sponge spicules—usually simple cylindrical rods though branching heads of tetractinellid spicules are sometimes represented. Within the cavity a very fine rod of silica, which must once have infilled the axial canal of the spicule, is often preserved;

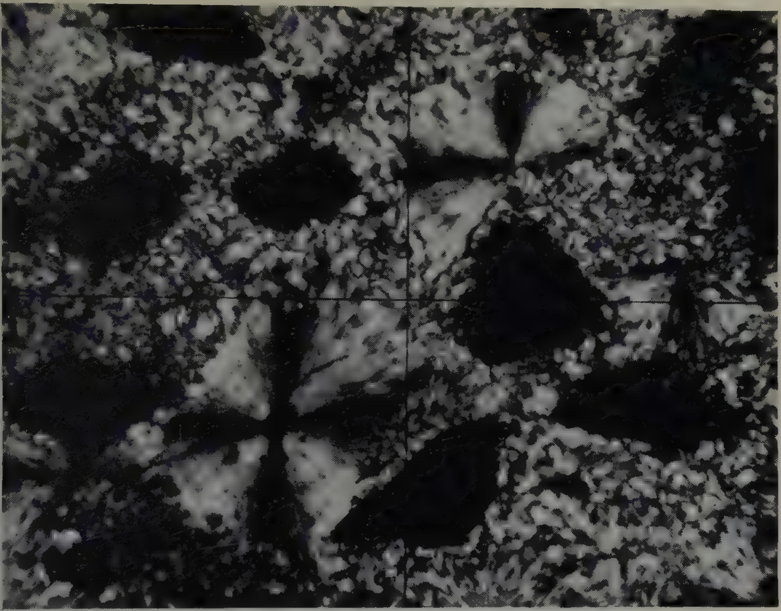
(c) Cavities of irregular shape produced by the dissolving out of small fragments of calcareous shells.

The relative importance of these three types varies considerably from locality to locality and in individual nodules. Normally some calcareous shell debris is still present, but in other cases all such material has been dissolved out. Limonitic oxides may be completely lacking, giving a pure white marginal rock, or may be so abundant as to stain it yellow, buff or brown.

a. Microscopic Features

In thin section under the microscope, the cherts are seen to be relatively pure siliceous rocks. 'Floating' detrital grains of quartz or glauconite occur, but a careful search of the rock slice is often needed to reveal their presence. It is clear therefore that the cherts do not result from simple silica-cementation of the surrounding sediment. The main characteristics of the chert core may now be summarised:

(*i*) *Lack of Detrital Grains*. Humphries (1956) found that the siliceous stone of the Lower Greensand showed a complete gradation from chert through sandy chert and cherty sandstone to a true sandstone. No corresponding sequence may be traced in the Upper Greensand where detrital



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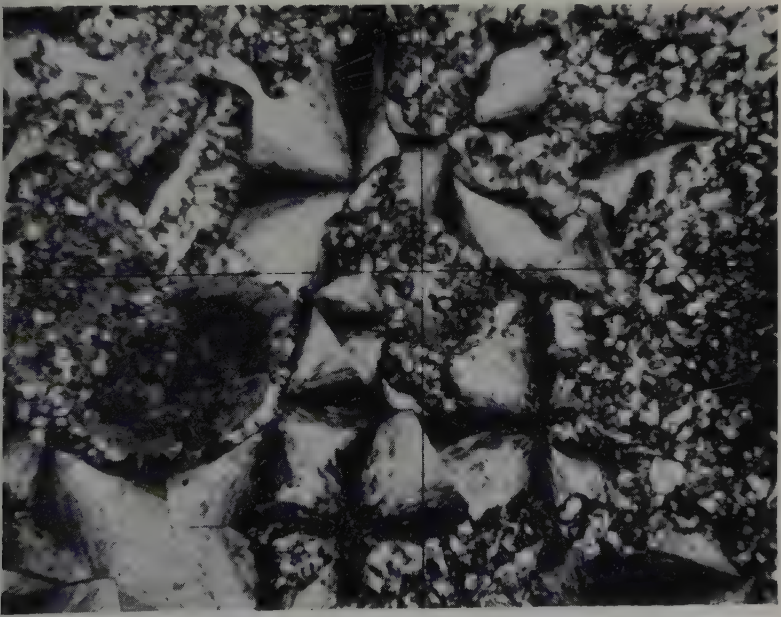

B

[*To face p.* 340

material (quartz and glauconite grains of fine sand or silt grade) is sparse in the great majority of chert nodules. Quartz occurs as rather angular grains which, in the rare cases where the groundmass is quartzitic rather than chalcedonic, may show secondary growth in optical continuity. In contrast, the grains of glauconite are usually rounded or botryoidal in form, like those of the surrounding sediment. The mineral may occur as perfectly fresh dark green grains, or may show surface dissociation to a pale green powder, or oxidation to brown limonitic material.

The point that the siliceous sandstone found in the Blackdown Hills and the Vale of Wardour probably results from the secondary silicification of original calcareous concretions has already been made. However, the Upper Greensand of Cheddington (31/492062), north of Beaminster, contains both calcareous stone and what may fairly be described as cherty sandstone. The latter shows a pronounced parallel banding, with layers in which detrital grains are closely packed alternating with bands of purer chert with scattered detrital grains. The detrital material is chiefly quartz and glauconite grains of fine sand grade, but the presence of laths of both muscovite and biotite is an unusual feature.

(*ii*) *The Composition of the Groundmass.* With the diaphragm of the microscope closed to cut the light transmitted to a minimum and so emphasise the phase contrasts, the bulk of the groundmass is seen to consist of botryoidal masses of silica, similar to those discernible in the marginal rock. These masses were termed 'Colloform' by Rogers (1917) who identified them as resulting from the contraction, during solidification, of a mass of silica gel. They are composed of a mass of very fine fibres of chalcedonic silica, normally of completely random orientation. It is quite obvious that any plane cutting a random aggregate of fibres, will show far more transverse or near-transverse than longitudinal sections, and these innumerable cross-sections tend to give the aggregate a granular appearance. Where a suitable nucleus (a sand grain or shell fragment) is present, however, a thin acicular sheath has often grown out perpendicular to its surface (Plate 14 B) revealing the true nature of this material. Such chalcedonic aggregates were termed 'Metacolloidal Silica' by Rogers (1917).

Between the colloform masses of metacolloidal silica, the former cavities, produced by contraction of the original gel, are themselves infilled with acicular chalcedony—the 'Fibro-chalcedony' of Smith (1960). The needles, which show straight extinction, radiate out from a centre on a lobe of the enclosing botryoidal mass, and Plate 14 A shows the partial and complete extinction crosses produced by these fans under crossed nicols. Where more than one fan infills a single cavity, the junctions of the fans are straight lines, giving rise to Smith's 'Polygonal Spherulites'.

The acicular chalcedony often appears brown in thin section, and the colour difference between it and the surrounding colloform masses may be very pronounced. This brown coloration does not result from iron-staining since under reflected light these areas appear milky white and opaque. Folk & Weaver (1952) and Pelto (1956) have shown that the coloration results from the presence of water in sub-microscopic cavities between the fibres which constitute the chalcedony. This enclosed water lowers the refractive index of the mineral well below that of quartz. (The chalcedonic fibres of the metacolloidal aggregates also show refractive indices below those of the included quartz grains.) The interstitial fans of chalcedony appear to have been deposited from normal aqueous solution and show no features indicative of a colloidal origin. In rare instances the fans of chalcedony do not completely infill the 'colloform cavity', the centre of which then contains a granular mosaic of authigenic quartz.

Occasionally the contraction of the silica gel mass has caused a complete rupture between it and the surrounding sediment. A brown chert from Durdle Door (30/803803) shows such a cavity, now infilled with acicular chalcedony which has grown out from either side, sometimes with a thin median band of granular quartz. A grey chert from Haven Cliff, Seaton (30/257897), shows a quartz vein, a quarter-inch in width, infilling a similar cavity, while smaller tension gashes, also quartz-infilled, are seen in the outer part of the chert nodule itself.

The cherts sometimes contain minute globules of opal, which were deduced by Smith (1960, 7) to be 'adventitious inclusions analogous to entombed sand-grains and glauconite'. These are, however, rare, and opal appears to be completely lacking from the groundmass itself. The colloidal globules mentioned by Hill (Jukes-Browne, 1900) were not seen, though it is evident that the colloform masses represent his chalcedonised globules.

(*iii*) *Organic Material.* Siliceous sponge spicules are invariably present and in the purest cherts may be the predominant material included. None of the Upper Greensand cherts examined could, however, be regarded as sponge-rocks of the type envisaged by Hinde (1885). The spicule walls are never preserved in their original amorphous state, the opaline silica having invariably been replaced by chalcedonic silica like that forming the colloform masses. The former axial canals are infilled by chalcedony, glauconite or brown ferruginous material; the canal has often been enlarged by solution, and a continuation of this process may eventually lead to the entire spicule being replaced by glauconite or limonite.

It is difficult to generalise on the occurrence of shell fragments within the cherts. In many cases they must once have been more abundant than were detrital grains, but whereas the latter are preserved unchanged, included calcitic remains are very prone to silicification. Even when completely replaced, however, they can still be recognised as dark patches of silica, so finely grained as to show 'pin-point birefringence'. These patches correspond in general size and shape to the shell fragments, and sometimes distinctive organic forms (chambered foraminifera and fragments of echinoid test and the mesh of holothurians) are still identifiable. In addition residual patches of calcite are sometimes preserved within the silicified remains.

There is no exact correlation between the state of calcitic remains in the chert and the lithological nature of the surrounding sediment. Where the host has been leached free from calcareous material, that present in the chert has either been dissolved out or completely silicified. However, within a calcareous sediment, the cherts may contain organic remains as unreplaced calcite or in a partially or wholly silicified state, the degree of replacement varying from bed to bed or even within a single nodule. Plate 14 B shows completely replaced and partially silicified shell fragments in juxtaposition.

Briefly, then, in the true cherts and their marginal rocks detrital grains are sparse. Calcareous material is of more importance; in the core it may be preserved unchanged, or may be partially or wholly replaced by chalcedony. Calcareous material in the marginal rock may be in a similar state to that in the core, but any unaltered calcite has usually been dissolved out to leave solution cavities in the chalcedonic groundmass. Sponge spicules are almost invariably present, though forming an unimportant part of the whole. Similar included material, detrital and organic, is usually present in both core and marginal rock. Botryoidal (colloform) masses are common to both core and marginal rock, and suggest that the chert originated as a gel. In the core the former cavities between these masses are themselves infilled with chalcedony, apparently deposited from aqueous solution. Such interstitial sheaths of chalcedony may be completely lacking from the marginal rock, but normally fill at least some of the cavities here. The chief distinction between chert and marginal rock is the more porous nature of the latter. Some of the cavities are original, formed as the gel mass solidified and contracted; the remainder result from the dissolving out of spicules or shell debris.

5. THEORIES OF CHERT FORMATION

The lack of agreement on the basic fundamentals of chert formation and the contradictions to be found in the literature have already been noted. Though the vast majority of the papers produced have concerned themselves exclusively with cherts occurring in beds of limestone, similar controversies have arisen over the cherts present in the Greensands of southern England. Three theories of formation have been put forward to explain these cherts:

(a) Redistribution of organic silica within the sediment—proposed by Hinde (1885).

(b) Silicification, by percolating solutions, of limestone doggers included in the sediment—proposed by Richardson (1947).

(c) Electrolytic precipitation of masses of silica gel on the sea-floor—proposed by Humphries (1956).

(a) Replacement Theory

Of these three theories, the second may be readily eliminated as far as the Upper Greensand is concerned, despite Richardson's convincing demonstration that the cherts of the Aptian Bargate Beds had been so formed. He showed that a complete transition could be traced from the calcareous Bargate stone, through semi-silicified types to a true chert. An organic source for the silica was rejected, and electrolytic precipitation of silica gel in a shallow sea was the source proposed. In the post-depositional period, limestone doggers were first formed. Subsequently the dispersed gel was dissolved and the doggers silicified to a degree dependent on the amount of silica originally included.

There are, however, no grounds for assuming that the cherts of the Upper Greensand originated in this way. It is true that in most areas (the Blackdown Hills being a notable exception) beds with chert also contain nodules of limestone but there is no trace of any transition between the two. Sometimes, as at Mere (31/803323), the same bed contains concretions of pure crystalline limestone, and nodules of pure black chert. The former may contain calcite-replaced sponge spicules, the latter silicified shell debris, but these form an unimportant part of the whole and other signs of replacement are totally lacking. Elsewhere, most notably on the Devon coast, a stratified arrangement is characteristic. Chert occurs in discontinuous bands, separated by beds of calcareous sands, often containing doggers of limestone which tend to be aligned in definite bands. Such an arrangement suggests a segregational origin but does not support the view that the chert nodules formed by replacement of limestone.

Replacement of the country rock may also be ruled out, since the quartz and glauconite which compose the Upper Greensand are both notably insoluble minerals. Dapples (1959) has described how, in the late stages of diagenesis under conditions of deep burial, quartz may be replaced by calcite. However, the high temperatures and pressures which result in the increased solubility of quartz also raise the solubility of amorphous silica. No conditions can be postulated under which the highly insoluble mineral silica could be replaced by the relatively stable amorphous type.

It is thus safe to conclude that, though replacement may have contributed to the formation of the Upper Greensand cherts, it was not here the dominant process that it was in the case of the Bargate cherts. The theories of Hinde and Humphries remain as possibilities and may now be considered in turn.

(b) The 'Sponge-bed' Theory

Hinde in his classic paper of 1885 concluded that the cherts present in both the Lower and the Upper Greensand were so dominantly composed of sponge spicules as to be classifiable as true organic deposits. He believed that they originated as 'spicule-mats', so densely packed that detrital grains were almost completely excluded, which were converted into true chert by solution and reprecipitation of the spicular silica.

The evidence on which this theory was based seems to have been essentially circumstantial—particular emphasis being placed on the general association of spicules and chert. While it is true that spicules are almost invariably present in the cherts of the Upper Greensand, they normally occur in relatively limited numbers, and even those cherts richest in spicules are far from being felted mats of the type envisaged by Hinde. In addition, the groundmass of both chert and marginal rock is composed not of partially dissolved spicules but of the colloform masses of silica indicative of an original gel.

The marginal rocks of the chert nodules, with their empty moulds of former spicules, Hinde described as follows: 'Thus by a peculiar sequence of conditions, the spicules in this portion of the rock have first been embedded in a siliceous matrix, then their hollow canals have been infilled with silica, and finally the spicules themselves have been completely dissolved away, leaving only the casts of their outer form and their solidified canals.' This accurately describes the typical marginal rock of the Upper Greensand cherts and would seem to show quite conclusively that the siliceous groundmass formed prior to the solution of the spicules. However, Hinde suggested elsewhere that it was the reprecipitation of the spicular silica here dissolved which produced the chalcedonic groundmass of the core. This apparent inconsistency, which was never resolved, is one of the major weaknesses of Hinde's theory.

(c) The Silica Gel Theory

The work of Humphries (1956) on the cherts of the Hythe Beds of the western Weald led him to the conclusion that the part played by spicules was a quite insignificant one. Since the cherts were involved in small-scale slumps within the sediment, he concluded that they must have been of contemporaneous origin. The colloform structures shown by the chert, together with the 'floating' detrital grains, were thought to point conclusively to an original gel. Precipitation might be caused by the electrolytes present in sea-water or by the action of algae or bacteria. Still water was essential and Humphries envisaged precipitation of the silica gel within masses of sea-weed or sponge-colonies so dense as to screen out detrital grains almost completely.

Despite the considerable inherent chemical difficulties, the precipitation of a mass of silica gel on the sea-bed is the only process which will explain all the following features, which are typical of Upper Greensand cherts:

(1) The colloform structures of both chert and marginal rock;
(2) The metacolloidal nature of the silica comprising these colloform masses;
(3) The general lack of detrital grains;
(4) The absence of spicules in the numbers required by Hinde;
(5) The perfect preservation of spicule casts in the groundmass of the marginal rock.

The formation of botryoidal and spherulitic structures depends on the fact that colloids show the properties of liquids rather than solids and so have a natural tendency to contract into spherical drops. The term 'Colloform' was introduced in 1917 by Rogers for the 'rounded, more or less spherical forms assumed by colloidal and metacolloidal substances in open spaces'. Taliaferro (1934 and 1935) identified opals and pure cherts as hardened silica gels which had contracted and dehydrated under natural conditions just as did silica gels in the laboratory.

Equally significant is the state of the silica composing the colloform masses. In thin section under crossed nicols, it is seen to consist of an aggregate of very fine fibres of chalcedony, usually of quite random orientation, though individual spherules may show a rudimentary radiate growth. Rogers employed the term 'Metacolloidal' for this disordered microcrystalline structure which is characteristic of solidified gels. In the laboratory, silica gels invariably solidify as the unstable amorphous form which is, in the course of time, converted to the metacolloidal chalcedonic state. Though it is possible that chert, solidifying under natural conditions, shows a similar change, it seems more likely that it would produce the stable chalcedonic form direct. The time factor operating in geological processes can never be reproduced in the laboratory, and the immense slowness of geological change may well give final results which cannot be artificially duplicated.

The reasons for rejecting Hinde's view of an original 'spicule-mat' have already been stated. In particular the perfect preservation of spicule moulds in the groundmass of the marginal rock does not suggest that this groundmass can itself have formed from dissolved spicular silica. The relative purity of the cherts and especially the lack of detrital grains is also worthy of note. Davis (1918) showed that silica gel, precipitated from a solution charged with fine sediment in suspension, was able to expel these impurities to form lenses of pure silica.

There is thus a considerable body of evidence to support the views put forward by Humphries. Unfortunately there is also overwhelming chemical evidence against the theory, at least in the form outlined above. Humphries, though puzzled by the lack of a suitable precipitant for silica at natural concentrations, believed that, if a sufficient concentration of silica could be built up, electrolytic precipitation of the gel would occur. This view was based on the assumption that silica in natural waters was present in the colloidal state. Chemical studies over the past thirty-five years have, however, demonstrated conclusively that silica is in fact carried in true solution —probably as monosilicic acid, H_4SiO_4. Electrolytic precipitation as silica gel is then virtually impossible since, long before a sufficient concentration of silica could be built up, precipitation of the highly insoluble calcium and magnesium silicates would have occurred by reaction with the salts dissolved in sea-water.

The problems raised by the nature of silica in solution and the conditions of its precipitation have given rise to so many contradictions over the years, that a brief summary of our present state of knowledge may be of value.

(d) Silica in Solution

Most of the geological literature produced on this subject during the first half of the twentieth century propagated the view that silica was carried not in true solution but as a colloid. This assumption was based on the experiments of Kahlenberg & Lincoln (1898) whose results, obtained with prepared solutions and not natural waters, have been completely disproved by subsequent more detailed work.

The development of the colorimetric silico-molybdate method now provides a simple means of determining silica solubility which has been employed by many workers. The great merit of the method is that it will determine only the silica present in true solution, the colloidal form causing no reaction. In this way it can be shown that the silica carried in true solution in natural waters corresponds with the total silica determined by gravimetric analysis, so that colloidal silica is not present. It was demonstrated by Alexander, Heston & Iler (1954) that silica in solution was hydrated to the monosilicic acid H_4SiO_4.

It follows that the experimental results of such workers as Lovering (1923) and Moore & Maynard (1929) who succeeded in precipitating silica gel from artificial colloidal solutions by the action of electrolytes, cannot apply to natural waters for two reasons. The process requires solutions of a concentration never attained by natural waters (more than 1000 p.p.m. according to Lovering) and the presence of silica in true solution would lead to the precipitation of the insoluble calcium and magnesium silicates at concentrations far below those required to produce silica gel.

Thus since the cherts of the Upper Greensand, and also of many other formations, show clear evidence of having formed from an original gel, some alternative means of precipitation must be found. Davis (1918) claimed that organic agents were responsible and showed experimentally that ammonium carbonate precipitated silica gel from a solution of sodium silicate. Sargent (1923) and Lees (1929) supported the view that ammonium salts, produced by the decomposition of organic remains, caused the precipitation of silica gel. An alternative view has been forwarded by Siever (1957), Pittman (1959) and Krauskopf (1959) who completely rule out precipitation of an original gel and claim that the transformation of silica from its original soluble state to its final incorporation in a chert nodule occurs only via the skeletons of siliceous organisms—a process exactly analogous to that postulated by Hinde. This view does not, however explain the colloform structures and metacolloidal nature of the silica both of which definitely imply an original colloidal state.

Since in Britain chert formation occurred most extensively during th Carboniferous and Cretaceous periods, it is not unreasonable to assum that conditions were then particularly favourable in that natural water were more than usually rich in silica. A warm climate would be an obviou attribute, and it may here be noted that Carter (1871) studied the spicule of the Haldon Greensand and deduced that the sponges lived in fairly shallow waters in a climate considerably warmer than at present. Low-lyin lands with a heavy vegetation cover would favour chemical rather tha mechanical weathering. Under such conditions, the waters would b silica-enriched, with a number of possible results:

(1) Siliceous organisms such as diatoms and radiolaria could flourish and prolific sponge growth might also be expected, though there is littl evidence that sponge colonies in the present seas are so governed.

(2) Some of the silica in solution would certainly be utilised in the formation of the glauconite so characteristic of the 'Greensand'. Bien *et al.* (1959 studied the effects of the mixing of river- and sea-water at the mouth of th Mississippi, and concluded that the marked decrease in silica was no wholly the result of dilution but also depended on its adsorption on t colloidal material and suspended solids in the river waters, as these cam into contact with the electrolytes of sea-water. Presumably a similar proces would operate in Cretaceous seas. Much of the silica-enriched materia would be carried out to sea and incorporated into the Gault Clay. Th remainder would flocculate into small masses which would be deposite among the sand grains of the sea-floor and there converted, by adsorptio of potassium and other ions from sea-water, into glauconite. The chemica weathering of the granitic masses of south-west England, and the halmyrolitic weathering of granitic detritus, would give waters rich in both silic and potash, and so favour this transformation.

(3) Silica might also be precipitated in the form of a gel to give rise to th chert nodules. From the arguments summarised above it is clear that sea

water electrolytes cannot have been responsible but that ammonium salts, produced by organic decay, are possible agents. Sponge beds would then provide an environment eminently suited to gel formation, which would account both for the incorporated spicules and for the often localised nature of the precipitation (a factor not readily reconciled with the electrolytic theory). The occurrence of chert in bands reflects the build-up, following each successive precipitation, of silica, or the ammonium salts, or both, to the level required for precipitation. This would be unlikely to occur if there were strong bottom currents and tranquil conditions would thus favour chert formation.

Davis (1918), in his experimental demonstration of this process, found that pure silica gel could be precipitated from a solution containing fine detritus in suspension—the flocculating gel expelling these impurities. This could account for the purity of the chert nodules, without resorting to the impenetrable beds of seaweed or sponges invoked by Humphries. The presence of shell fragments within the chert is significant in this respect; larger and heavier than the fine sand and silt grains of the surrounding sediment, they would be less readily expelled from the gel.

Masses of relatively pure silica gel so precipitated on the sea-bed are believed to have given rise to the cherts of the Upper Greensand, though not by simple dehydration and contraction. The conversion of the hardened gel into true chert is thought to depend on the diagenetic redistribution of the opaline silica of spicules included in the sediment. Thus it is contended that the silica of the chert nodules is derived partly from inorganic and partly from organic sources.

Evidence in support of this view is provided by both the chert cores and their marginal rocks. The latter are characterised by cavities produced by the solution of sponge spicules. The perfect preservation of the form of the spicules, even to the fine rod which infilled the axial canal, proves that (a) contraction and solidification of the gel were completed prior to the solution, and (b) at the time of solution the groundmass was in its present, relatively insoluble, chalcedonic state while the spicules retained their soluble opaline form. Solution must thus have occurred at an early stage before the unstable opaline silica was converted on 'ageing' to the stable metacolloidal form.

While Hinde (1885) suggested that the silica so dissolved was reprecipitated in the chert core, Humphries (1956) believed it to be carried out of the sediment in solution, so making no contribution to the chert. This would imply that the arresting of the process when only the outer zone of the chert had been affected was purely fortuitous. However, similar rocks surround the cherts of unconsolidated sands, which would afford a free passage to the solutions, and those of fine silty sediments—such as occur at Mere (31/803323) and Chard (31/313088)—where percolation of water would have been more restricted. This constancy of association cannot be lightly dismissed as merely coincidental.

Moreover, if Humphries's view is correct, the marginal rock should consist of a solidified gel with casts of spicules, the core of a solidified gel with undissolved spicules. This is not in fact the case. While much of the groundmass shows the colloform structures and metacolloidal state indicative of a gelatinous origin, not all the silica present is of this type. This is only to be expected, since the colloform structures result from the contraction of a gel during its dehydration. In this way, cavities are produced between the botryoidal growths of silica, and if these spaces are themselves silica-infilled, this is obviously a subsequent, secondary development. In fact the cavities are infilled with fans of acicular chalcedony, and occasional mosaics of granular quartz, which appear to have been deposited from normal aqueous solution, and show no indications of a former colloidal state.

In both the core and the marginal rock, an original mass of colloidal silica was converted, by dehydration and contraction, into a solid rock, so producing the colloform groundmass. Subsequently the opaline silica of incorporated sponge spicules was taken into solution in the marginal rock, leaving empty spicule casts preserved in the groundmass, and chalcedonic silica was deposited from solution in the core, converting the porous rock into a true chert. Though there is no absolute proof that the silica deposited from solution in the core was the same silica taken into solution in the marginal rock, the constancy of the association of two quite different rock types in a single nodule leaves no reasonable grounds for doubting that this was in fact the case.

It was noted in an earlier section that chert and limestone tend to occur in alternating bands within the Upper Greensand. Though signs of replacement are generally lacking, silicified shell debris may occur in the cherts, and calcite-replaced spicules in the limestone. Thus converse effects of solution and reprecipitation must occur within a few feet, sometimes even a few inches, of each other. This seemingly improbable process may in fact be accomplished by relatively slight variations of pH within the sediment, since the solubility of calcium carbonate is very susceptible to such changes. The figures here given are those of Correns (quoted by Newell *et al.*, 1953), the solubilities being expressed in parts per million of sea-water:

Under slightly acid conditions	(pH 6.5):	500 p.p.m. $CaCO_3$, 240 p.p.m. SiO_2.
neutral	(pH 7):	250 p.p.m. $CaCO_3$, 250 p.p.m. SiO_2.
slightly alkaline	(pH 7.5):	90 p.p.m. $CaCO_3$, 265 p.p.m. SiO_2.

If these figures are correct, it is clear that under acid conditions, calcite is more soluble than amorphous silica, but that under alkaline conditions amorphous silica is the more soluble of the two. Since chemical replacement depends on a soluble substance being in contact with a solution charged with one less soluble, it follows that under acid conditions, calcite will be replaced by silica, while under alkaline conditions, silica will be replaced by calcite.

The intimate association of decomposing organic matter with the masses of silica gel may tend to lower their pH below that of the surrounding sediment. The slightly acid conditions will then favour the deposition of silica within the chert bands and will lead to a dearth of silica in the connate waters here. Diffusion of silica from adjoining areas will tend to restore the equilibrium but will lead to the waters in these latter areas being themselves impoverished in silica. Silica will therefore enter into solution where it is available in a suitable form (e.g. the opaline silica of the spicules) until the silica content of the waters is again built up—whereupon further deposition will occur within the chert to recommence the cycle. A corresponding 'outward' migration of calcareous material may result in the formation of doggers of limestone or calcareous sandstone in the sands separating the chert bands.

Though some such two-way migration system is undoubtedly set up, the actual mechanics of the process are still uncertain. Newell *et al.* (1953), following Correns, suggested that replacement of limestone by chert was governed by slight variations of pH; Smith (1959) reiterated this view.

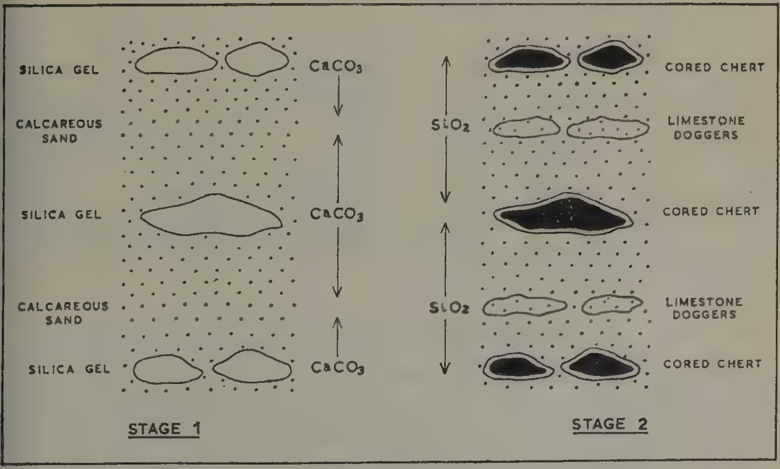

Fig. 4

However, Krauskopf (1959) challenged Corren's statement that the solubility of amorphous silica increases with rise of pH, and claimed that, at the alkalinities found in natural waters (which seldom exceed pH 9) silica is no more soluble than in acid solution. He does, however, demonstrate an increase in silica solubility with rising temperature. Dapples (1959, 50) makes a similar point: 'Conceivably temperature may be the important control and centres of intense organic activity might produce loci of temperature sufficiently high to dissolve silica, whereas in the adjacent positions such interstitial waters would be oversaturated and deposit silica.' Whether controlled by pH or temperature or a combination of the two, the net result of the process was to cause differentiation of the soluble components of the sediment, as shown in diagrammatic form in Fig. 4.

6. CONCLUSIONS

The evidence shown by the Upper Greensand cherts suggests that they have formed from masses of silica gel precipitated on the sea-bed. The porous rock produced as the gel solidifies is converted into a true chert by the deposition of silica from solution. This silica is derived from the solution of opaline sponge spicules in the marginal zone of the chert nodule or the surrounding sands. The solution and reprecipitation of the silica is thought to depend on slight variations of acidity, or possible of temperature, within the sediment. This process is believed to provide a more complete explanation of the features shown by the cherts than do any of the proposed alternatives.

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EXPLANATION OF PLATE 14

A. Grey chert from Mancroft (31/272088), three miles west of Chard, Somerset. (Under x.n., ×80)

The groundmass consists of a random aggregate of fine chalcedonic fibres, with acicular chalcedony infilling former cavities and giving the prominent extinction crosses. Detrital grains are lacking and the black patches are cavities produced by solution of the included calcareous material. The dark patch in the extreme north-east is a silicified shell fragment showing the characteristic 'pin-point' birefringence

3. Chert from Snowdon Hill quarry (31/313088) Chard, Somerset. (Under x.n., ×60) Fans of acicular chalcedony, producing incomplete extinction crosses, infill cavities between colloform masses of metacolloidal silica. In the west of the field a calcareous shell fragment shows only incipient silicification, while another, immediately to the north, has been completely silicified. Both are surrounded by thin sheaths of acicular chalcedony which have grown out perpendicular to their surfaces

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DISCUSSION

DR. W. E. SMITH congratulated the author on his study of a very neglected group of rocks. He stated that the Albian chert within Wessex displays considerable variations in appearance, microstructure and relationship with the host-rocks. For example, core-chert is typically developed only in the lower part of the chert beds near Seaton; a higher levels the translucent chert makes sharp contacts with the enveloping calcarenite. The core-cherts are typically black and rarely contain geodes whereas the chert in the higher levels is typically yellowish and translucent and often contain geodes lined with mammillated chalcedony or euhedral quartz. Furthermore, in the upper part of the Upper Greensand in the Seaton and Beer districts chert occurs a transgressive sheets and irregular veins. Dr. Smith therefore asked whether the autho claimed that all the Albian cherts can be explained by the single mechanism that he ha outlined. If not, would he indicate more precisely the types of chert to which his conclusions apply. Dr. Smith did not accept the author's claim that there is very littl difference between the crust and core of core-chert, apart from porosity. The infillin of the pores with dark lubricating oil does not reproduce the appearance of the cores Detailed microscopic examination by reflected light shows that the crusts contain mor or less uniformly dispersed, minute white particles believed to be clay minerals. Thes are very scarce in the core although concentrated at the interface as a narrow whit band, which is often visible in hand-specimens. The core therefore appears to be clarified zone from which the clay minerals have migrated, probably during the ge stage. Can the author account for this contrast between core and crust?

Silica gel deposited from dilute aqueous sols has a very low viscosity and rapidl spreads out into a flat-topped sheet. The shapes of most of the chert masses are quit inconsistent with the primary deposition of such gel on the sea-floor. Sheet-like tabula chert does occur but often swells out into beaded or bulbous masses. Furthermore irregularly shaped nodular masses several feet high and with their greatest diamete normal to the bedding are not uncommon. It is difficult to imagine how these masse could have retained their shape if deposited as gel in the sea-floor. Dr. Smith was i complete agreement with the author as to the minor role played by sponge spicules, bu considered that he has greatly underestimated the importance of post-depositional o diagenetic segregation of inorganic silica gel. Silica gel could have been irregularl deposited with calcareous ooze and segregated during diagenesis to give the embryoni chert masses. The irregular shapes of these gel masses can then be explained by the buttressing action of the enclosing sediment and the more viscous character of the ge that could be deposited from connate waters. The growth of the embryonic gel masses could be due to dilation or replacement by such a mechanism as outlined by Correns

MR. DRUMMOND also congratulated Dr. Tresise on his very clear account, but disagree with his hypothesis that the Upper Greensand cherts are primarily formed from silic gel masses on the Cretaceous sea-floor.

In Devon, most cherts contain sedimentary structures such as bedding and false bedding, and in some cases these can be traced from the chert into the surrounding matrix, which in its turn is a coarse shelly calcarenite indicative of a shallow and turbulent environment.

The chert bed matrix is usually a highly calcareous rock with a low insoluble residue content, and the latter is dominantly of sub-silt size. This general absence of readil visible sand-grains within the matrix is the reason why such grains are usually absent from cherts, but where the matrix does become sandy, as at Maiden Newton, then the cherts themselves also contain sand-grains in some abundance. It is a feature of the Upper Greensand chert beds of Dorset, that as the sand content rises, not only are the chert beds less well developed, but the cherts themselves are impure.

At Henfords Marsh, near Warminster, a cherty sandstone occurs towards the top of the Upper Greensand. This is a resistant bed in a calcareous glauconitic sand, but has been completely decalcified itself to form a porous, very 'impure chert' or cherty sandstone. It seems, therefore, that as with the Lower Greensand chert beds, transitions from pure cherts to chertified matrix does in fact occur.

As pointed out by Dr. Tresise, with the exception of subsequently decalcified localities, cherts are invariably associated with limestones and highly calcareous rocks, and it is possible that they are a product of limestone diagenesis.

Deposition of a calcareous silt containing finely disseminated silica occurs under conditions of alkaline pH. With the addition of further layers of sediment, not only do the underlying beds begin to compact, but they moved beneath the sedimentary interface and into an environment of lower pH. Such changes within the top few feet of sediment may well provide the necessary chemical conditions to allow silica to migrate to the less alkaline environment where it becomes precipitated, and the Calcium Carbonate to migrate in the opposite direction, to produce the banded chert—limestone effect so characteristic of the chert beds.

Because the matrix of the chert beds are usually calcarenites with fine detritus, it is not necessary that colloidal silica as such should be a prerequisite of chert formation; abundant sub-silt sized quartz fragments, which because of their large surface area could be dissolved, may well provide the major source of silica.

Finally, I was interested to see that in one of the slides of chert there was a polygonal suture pattern preserved. This is very similar to that in the recrystallised calcareous matrix, and I would like to ask Dr. Tresise if he thinks this may be silica pseudomorphs of the original calcitic matrix.

In reply THE AUTHOR said that Dr. Smith correctly points out that not all Albian cherts are of the same type, but it should be emphasised that the restriction of cored types to the lower part of the chert beds applies only to the Seaton district. A few miles to the north, near Chard, cored nodules are characteristic of the highest chert-bearing beds.

The theory outlined by the author does not imply that all cherts are of the same type. It seems clear that diagenetic solution of spicules can occur either in the marginal part of the chert nodule or in the surrounding sediment; the former will give rise to cored cherts, the latter to the uniformly massive type.

The transgressive chert veins, which at Beer Harbour are found along the line of the periclinal axis postulated by Dr. Smith, do seem to be an exceptional case. They occur mmediately below the Chalk and in some cases are definitely associated with Cenomanian material infilling fractures in the top Upper Greensand. They are relatively rich in detrital grains and may show a gradational junction with the surrounding sediment. They thus appear to be replacive, and strictly speaking are probably Cenomanian rather than Albian in age.

Dr. Smith makes an interesting point concerning the differences between the core and the crust of the cherts, but his assumption that the core represents a zone of clarification and the crust a zone of accumulation would seem premature on this evidence. Where chert nodules occur in ferruginous sands, the marginal rock is often considerably stained by iron, simply because its porous nature renders it permeable to ground waters in a way which the core is not. Presumably materials other than iron oxides could be similarly introduced. Further research as to the exact nature of the material observed by Dr. Smith, and its possible origin, would seem to be essential before any conclusions are drawn.

The extent to which the form of the chert masses is controlled by (a) original deposition of silica gel or (b) migration of silica within the newly deposited sediment is obviously a matter of conjecture. The author would not deny the possibility of diagenetic migration of silica gel, provided this took place at a very early stage, when only displacement of the surrounding unconsolidated sediment is entailed.

Mr. Drummond's suggestion of actual replacement of a calcareous sediment seems much less tenable. It is true that the Upper Greensand is usually calcareous and is sometimes very richly so, but cherts are not confined to the most calcareous horizons. Nor can the almost total absence of detrital grains from the nodules be explained by replacement of the calcareous fraction of the surrounding sediment.

Where decalcification of the chert beds has occurred, the resulting decrease in volume has sometimes (at Hardown Hill and in the Axminster district) caused considerable disturbance in which all semblance of stratification has been lost. In other cases (the Blackdown Hills, the Vale of Wardour and near Westbury) there has been no perceptible disruption of the beds, and the original content of calcareous materia cannot have been very great. A common result of decalcification has been the conversion of the calcareous doggers present to siliceous sandstones. There is, however, no sign of any transition between this silicified stone and the true chert nodules which occur in calcareous and decalcified beds alike.

Comparison of the Albian chert with the flint found in the Cenomanian beds north of Le Havre is of interest in considering possible modes of origin. The Cenomanian of Normandy, with layers rich in sand and glauconite grains alternating with purer calcareous beds which contain bands of flint nodules and bands of limestone doggers looks remarkably like a deeper water facies of the Upper Greensand, and was indeed so regarded during the nineteenth century. In hand specimen the cored black flint bears a striking resemblance to some of the Albian cherts (most notably those of the Chard and Mere districts) whereas in thin section the features of the two are quite different. The colloform structures and acicular fans characteristic of the Upper Greensand types are totally lacking from the flint which shows irrefutable evidence of replacement, and often grades marginally into the surrounding chalk.

This complete contrast suggests that, even in those areas where conditions seem mos suited to a replacive origin, some other mode of formation must be sought for the Upper Greensand cherts. The theory outlined in the paper is believed to provide the most satisfactory explanation of the features (including the polygonal suture pattern mentioned by Mr. Drummond) shown by these cherts.

‘Basal Surface’ or ‘Weathering Front’

by J. A. MABBUTT

Received 17 March 1961

RUXTON AND BERRY (1959) have suggested ‘basal surface’ for the boundary between weathered and unweathered rock, in place of ‘basal platform’ as used by Linton (1955). They admit that this surface is ‘sometimes regular, sometimes intricately irregular, and at all times susceptible to ready generalisation’. For reasons stated below, the writer considers this term inapt, and proposes the term ‘weathering front’ to define the interface of fresh and weathered rock. It is noted that Büdel (1957) has used ‘Tiefenfront’ with this meaning.

Firstly, ‘weathering front’ is considered to give more adequate expression to the dynamic aspect of weathering than does ‘basal surface’, with its implication of an ultimate lower limit. The depth and zonal structure of the weathering profile normally change with time, and the weathering front advances until the profile ‘is in equilibrium with the surface environment’ (Ruxton & Berry, 1957). This mobility is implicit in the account by Ruxton & Berry (1959) in which they describe structural control of incipient weathering below the basal surface. Theoretically, the weathering front may rest in equilibrium at a mature stage in the weathering cycle, with associated landscape stability, but any renewal of erosion at the surface will bring rejuvenation of weathering and renewed movement of the weathering front. This is demonstrated by recent stream incision in Hong Kong (Berry & Ruxton, 1960).

Secondly, ‘basal’ may mislead in implying that weathering merely progresses downwards from a land surface. Apart from any lateral component induced by relief on the initiating land surface, weathering may be subject to strong structural compartmentation, particularly in massive rocks. Abrupt lateral changes from soft pallid zone to hard fresh rock are common, and the steeply rounded outlines of many granite domes exhumed from pre-weathered regoliths also indicate how irregular the weathering advance may be. Weathering front would appear to be preferable to ‘basal surface’ in that it is free from directional connotation.

Lastly, weathering may progress selectively and in depth, rather than massively on a single face, as suggested by ‘surface’. Core-stones and tor blocks are isolated and by-passed by the main advance; elsewhere, weathering may penetrate along narrow shear zones, far ahead of any general attainment of rock breakdown. This zonal nature of weathering attack is also comprehended in ‘weathering front’.

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A Geological Itinerary through the Spanish Pyrenees

(Written for the Summer Field Meeting, September 1961)

by J. M. RIOS

Received 26 June 1961

1. INTRODUCTION

GEOGRAPHICALLY the 'Pyrenees' denotes a mountain chain which extends from east to west, and separating Spain from France. The frontier approximately coincides with the line of the watershed. The Spanish Pyrenees are much more complex topographically than the French side, being wider and the descent from the axial line to the plains being completed more gradually across a series of parallel mountain chains. The French Pyrenees are much narrower, and consequently more precipitous.

The area sharing the same fundamental geological history with the Pyrenees, both in sedimentation and tectonics, extends much farther to the west, always with the same mountainous character, although gentler in

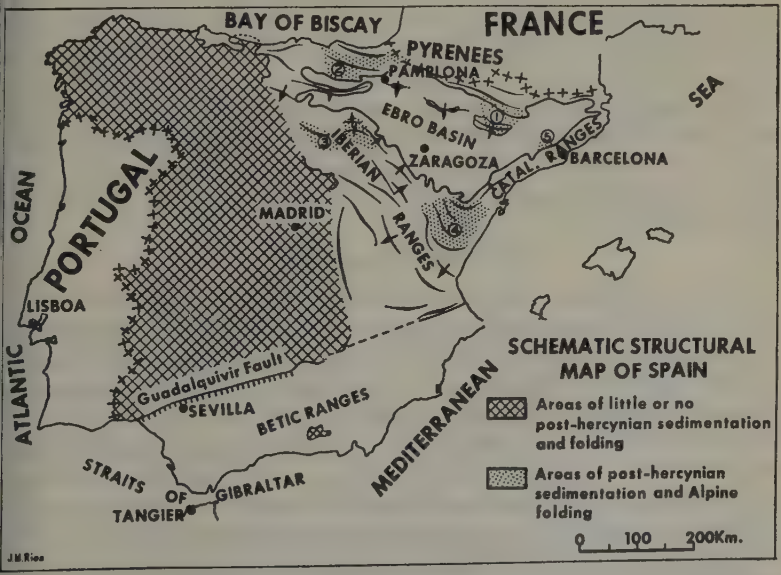

Fig. 1. Geological boundaries in Portugal omitted

some parts than in the Pyrenees proper. In this sense the Pyrenees extend well into the Asturian region along the Bay of Biscay, and include most of the ranges known as the Cantabrian Mountains.

The party will travel through the Pyrenees from east to west, just reaching the eastern part of the Cantabrian Mountains but leaving out the easternmost end of the chain. The route will cut across the different elements, from the Ebro Valley to the basement, several times. The structure of the chain is varied, complex but very well exposed. It is possible to follow clearly the change from simple to complicated series and back again to simple ones, and finally to see them develop into a very thick and complex succession. Most of the important changes affect the Cretaceous formations, and to a lesser degree the Eocene system. The route will also enable the tectonic evolution to be demonstrated.

Intense exploration for oil started recently and is yielding a rapid increase in our knowledge. Apart from the fact that many of the results are kept unpublished and that a detailed analysis would be too long, the following account aims to be only a general unified summary, uncomplicated by new discoveries still imperfectly digested.

2. GEOLOGICAL HISTORY OF THE IBERIAN PENINSULA

(a) Pre-Hercynian History

The pre-Hercynian history of the Peninsula appears to have been relatively uniform. A regime of marine sedimentation was dominant and on the whole fairly continuous from the Cambrian until well into the Upper Carboniferous.

Our present limited knowledge suggests that the Hercynian folding phases affected the whole of the area occupied by the present Peninsula in a similar manner. The Peninsula was during these times part of a much more extensive palaeo-geographical unit, an ancient continent which later was broken up.

(b) Post-Hercynian History

Following the close of the Hercynian orogeny at the end of the Carboniferous, the Peninsular Region became divided into two regions whose subsequent history was essentially different. The western region, bounded to the south by the Guadalquivir fault (or flexure) which originated at this time, and to the east by assumed flexures (hidden today under more recent sediments), shows from then on a continental development except for a few, short-lived marine invasions (mainly the Cretaceous transgression). The southern (south of the Guadalquivir fault) and eastern region (east of the flexures) began to function again as a geosyncline, and became the site of more or less thick, sometimes very thick, deposits, either of continental or marine character.

The complicated problems of the Betic Chains (to the south-east of the Guadalquivir fault) will not be considered here; it will suffice to say that their sedimentary development is varied and complex, although not essentially different from the rest of the peninsular areas of Alpine sedimentation. But their orogenic history is more prolonged (the folding lasts well into the Miocene) and also more violent. Among the Alpine fold systems in Spain, the Betic are by far the most complicated, even reaching a true Alpine tectonic category.

3. THE GEOLOGICAL SETTING OF THE PYRENEES

The Pyrenees can be defined geologically as those regions which make up the Pyrenaic orogen—that is the regions which, after the Hercynian folding, had the same geosynclinal history, and which together were subjected to a more or less uniform style of folding during the Alpine orogeny, building a continuous mountain chain. These folded sediments range from Permian to late Oligocene and rest on the Hercynian basement which is exposed almost continuously in the axial zone. This chain forms the Pyrenees proper and also its direct prolongations which share the same trends and display the same type of general structure.

The geological history of the Pyrenees is also shared to a varying degree by other chains (Iberian and Catalonian Coastal Ranges), but they are not true extensions but separate branches. The sediments of the Alpine geosyncline were deposited over a much wider area than the Pyrenees proper, extending to the south, for example, as far as the Ebro Valley where they are covered by Oligocene–Miocene sediments. The structure and history of the Pyrenees are tied up closely with that of the Ebro Valley so that one cannot separately study both regions.

4. GENERAL GEOGRAPHICAL AND GEOLOGICAL CHARACTER OF THE EBRO VALLEY

(a) Geographical Description

The Ebro Valley, in its present geographical expression, occupies the north-east corner of the Iberian Peninsula (Fig. 1). It has a triangular shape and is confined on the north by the Pyrenean chains along the French–Spanish border, on the east by the Catalonian Coastal Ranges, and on the south-west by the Iberian Mountain Ranges.

(b) Geological Description

The triangle formed by the Ebro Basin proper corresponds to a deep depression filled by Eocene and Oligocene sediments which, depending on heir more marginal or their more central location within the depression, appear today highly, little, or not at all folded, and are partially covered by a relatively thin sheet of Miocene deposits.

The mountain ranges that frame the basin are essentially built up by a more or less complete Mesozoic sedimentary sequence which is folded, and through which the Palaeozoic basement locally appears.

The Eocene–Oligocene formations, before they were largely removed by erosion, also spread over part of the area occupied by the framing mountain ranges and some isolated outcrops are still preserved. Similarly, there are some ramifications of the marginal Mesozoic chains of mountains which outcrop among the Eocene–Oligocene sediments of the Ebro Valley.

The three marginal chains are more or less of the same age, and originated during the Alpine phases which took place from late Eocene until the end of the Oligocene. The Pyrenean Cordillera is the most important and its tectonic category varies between that of the more violent Jurassic fold movements and that of the more moderate Alpine types; the Iberian Range is of an intermediate category and corresponds to the typical Jurassi folding; the Catalonian Coastal Range displays more moderate folding and in its constitution vertical movements played a more important par than folding.

The Palaeozoic basement displays intense folding which took plac during the Hercynian orogeny in Upper Carboniferous time.

Both in the Pyrenean and Catalonian Coastal Chains the Palaeozoi outcrops contain granitic intrusions which correspond to synorogeni Hercynian magmatic phases. Their regional metamorphic influence i small, at least compared with the metamorphic effects of the coeval graniti intrusions in the western half of the Peninsula which were very importan both in intensity and extent in the later geological history of the Peninsula

(c) Geological History of the Ebro Valley

(*i*) *Pre-Hercynian History and Hercynian Intrusions.* The present knowledge of the sedimentary and tectonic pre-Hercynian and Hercynia development of this region is both imperfect and incomplete. It has bee built up as a synthesis of a series of observations from outcrops which ar often discontinuous and distant from one another. All our knowledg points to it being essentially similar to that of the western half of th Peninsula. What, then, might be the reason for the difference in the post-Hercynian development?

Most probably it lies in the fact that the Hercynian orogeny in th Western Zone developed a very intense magmatic activity, predominantl of an acid granitic type, which caused a very intensive and extensive metamorphism. As a consequence the Western Zone became stabilised, incapable of further folding and tectonically dead, except in a regime o fracturing.

In the Eastern Zone, on the contrary, the magmatic phenomena must have been much less intense. Among the mountain ranges which frame the Ebro Valley, the Pyrenees show granitic Hercynian intrusions (Figs. 2 and 3) but their extent compared with that in the area occupied by the pre-Hercynian formations in the Western Zone of the Peninsula is much smaller and their metamorphic action much less.

The Catalonian Coastal Chain shows a more important Hercynian magmatic activity. The Iberian chains do not show any granitic outcrops, but it should be remembered that, generally speaking, the Mesozoic and Caenozoic cover is much more complete, the erosion surfaces here being at a higher stratigraphic level than in the Pyrenees.

(*ii*) *Post-Hercynian Mobility.* The conclusion that the Hercynian granitic intrusions were much less important in the eastern half than in the western half of the Peninsula, is more due to inference than direct observation. It is based on the fact that the eastern part of the Peninsula did not become immobilised as did the western one. On the contrary, it was able to function anew as an orogenic area, with prolonged and deep subsidence, the accumulation of new and sometimes very thick marine and continental deposits, and, finally, the creation of important systems of folding. It did not acquire the rigidity, which in the western part is due to strong and extensive metamorphism and crystallisation.

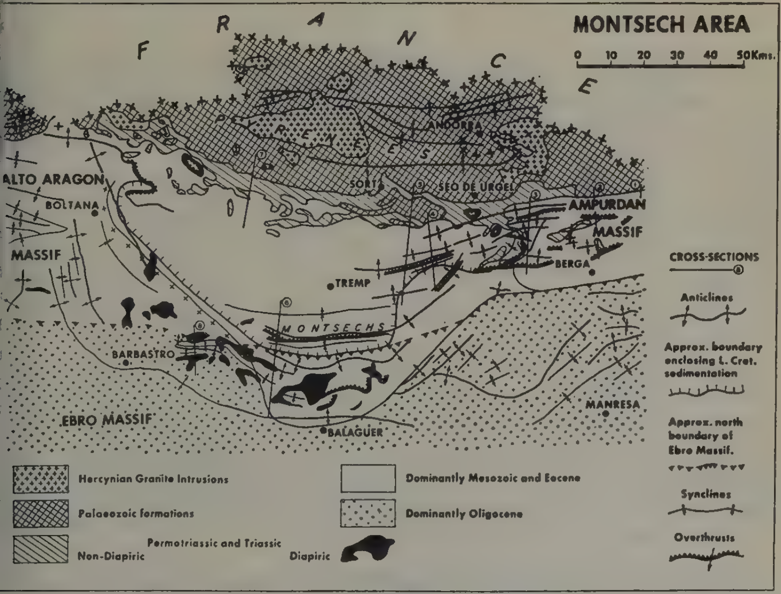

Fig. 2

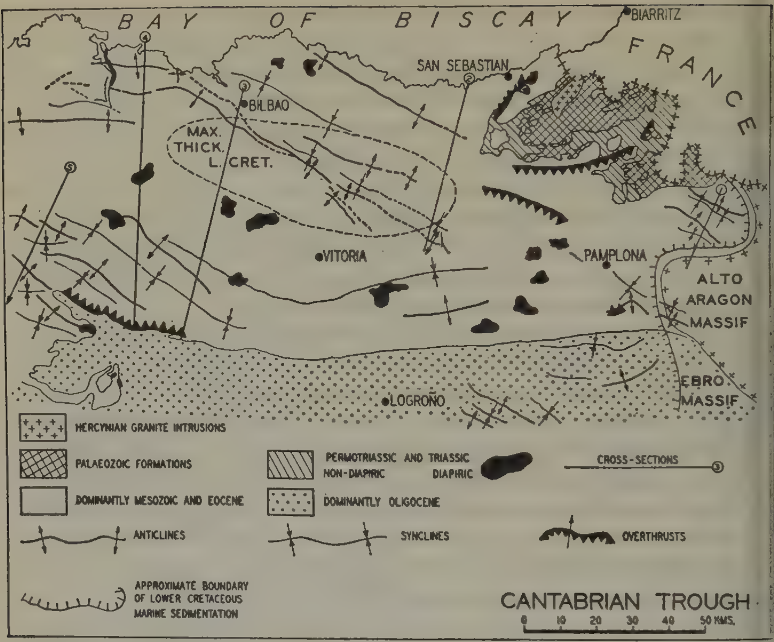

Fig. 3

(*iii*) *Triassic–Jurassic History* (*Uniformity*). If we examine the characte of the post-Hercynian formations which fill the Ebro Valley a salien fact strikes us: the sediments are patchy and variable, particularly abov the base of the Cretaceous.

After the Hercynian folding was completed and until the end of Jurassic the sediments of the Permian, Trias and Jurassic are uniform to a certai degree[1] in their respective lithologies, and somewhat less so in the thicknes of their units which is more variable from place to place. All throug this period there was relative uniformity in the depositional basins an no great regional variations.

(*iv*) *Cretaceous History* (*Mobility and Variety*). From the late Jurassic until the Middle Cretaceous, the depositional basins acquired a grea mobility, although the pattern was very variable. Some areas develope as great deeps and very thick and continuous sedimentary series wer deposited; others emerged and a greater or lesser part of their previou deposits fell prey to erosion; others oscillated vertically which gave ris to a very irregular and discontinuous series.

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[1] To a certain degree because compared to conditions in other areas, for example in Afric and America, all geologic conditions are extremely variable in Spain. This country is extraordinaril rich in quantity and variety of stratigraphic and tectonic situations, as a consequence of the rapidity of changes, Spain, in great areas, is a geological mosaic.

The stratigraphy of the Upper Jurassic–Middle Cretaceous is extremely varied and complex. Some series are complete and entirely marine; some are continental sediments, occasionally interrupted by one or more marine episodes. In other areas this was a period of erosion and a greater or lesser part of the Jurassic, and even Triassic and Permo-Triassic rocks disappeared, and the Palaeozoic basement was exposed.

In addition to the mobility of the basins from the end of the Jurassic onward and the variety in character of the sediments, there were also discordances and even the formation of basal conglomerates—all phenomena suggesting a mild phase of folding at this time.

It is difficult to discern whether these movements were due to vertical displacements of blocks within the Palaeozoic basement, or to its tectonic re-formation in ample folds; or perhaps, as a final cause, to deep magmatic movements. But it is a fact that after the end of the Jurassic all the area of the present Ebro Valley and marginal chains experienced a deep structural modification. The previous, rather uniform bottom became divided into compartments by a series of elevations separating small local deeps. These elevations of the basement, known in the Spanish geological literature as the 'Macizo Aragones', 'Macizo del Ampurdan', etc. ('Ebro Masse', 'Ampurdan Schwelle', etc., of the German geologists who first showed their existence), received very sparse and fragmentary sedimentation or none at all. The deeps received thick sedimentary deposits in complete and continuous series. There were also zones of intermediate characteristics.

(*v*) *Regions of deposition of thick and continuous Cretaceous sedimentation.* As can be seen in the map (Fig. 1), the special regions characterised by the accumulation of thick sediments are distributed at present along the margins of the Ebro Valley and are a part of the Sierras which frame the valley. These areas are as follows:

(a) **Montsech Area** is characterised above all by the extreme difference which exists between the complexity of its almost complete stratigraphic sequences and the simplicity of the neighbouring ones to the east, south and west where many of the members are lacking, and also by the great difference in thicknesses between formations within and outside the area. There is important diapirism all around the margins, especially on the south side. The diapiric material belongs to the Keuper. An unknown number of diapiric manifestations are hidden by Oligocene sediments.

(b) **Fosa Cantabrica** is characterised by the exceptional accumulation of Cretaceous sediments (both Lower and Upper) which contrasts with the neighbouring areas, especially to the east. This area is also characterised by its intense diapirism—the most important on Spanish soil and most of it of the salt dome type.

(c) **Sierra de la Demanda** is characterised by the thickness of its Lower Cretaceous with respect to that of surrounding zones, and also by its continental character. The Keuper around the area is hardly diapiric at all.

(d) **Maestrazgo Area** has more or less the same characteristics but, in the Lower Cretaceous, marine facies are represented along with the continental ones.

(e) **Montserrat.** This area is not a part of the marginal chains but belongs to the Ebro Valley itself, and typifies some of the facies changes which occur there. It is altogether different and much less important than the previous areas, both in extent and in thicknesses involved. It is a typical example of variability in the Palaeocene inside the Ebro Valley proper, otherwise so monotonous in all its characteristics. There are many more areas of the same character along the inner border of the marginal chains.

Of these regions only (a) and (b) will be included in our visit.

(*vi*) *Geological History of Post-Cretaceous Time*. The deposits of the Albian already show a trend to a greater uniformity of character in the sediments of most regions. Immediately afterwards there was a great Cretaceous transgression which began in the Cenomanian and persisted until the end of the Senonian. The Upper Cretaceous sediments are rather uniform in their lithological character in all the marginal areas of the Ebro Valley, but less so in their thicknesses.

The transition from Cretaceous to Eocene was marked by a general withdrawal of the seas which became permanent for the lower or southern half of the Ebro Valley, south of a line which constitutes the median line of the triangle going from the north-west corner to the middle point of the Catalonian Coastal Range. North of this line the sedimentation soon reverted to marine conditions, whereas to the south of it the continental regime persisted. The thickness of the Eocene sediments is great in both facies, and also variable, especially in the marine examples.

At the end of the Eocene began the great Alpine folding which was to create the mountain chains which today frame the Ebro Valley. The rise of the Catalonian Coastal Range isolated the Ebro Valley from now on from all communication with the open Mediterranean Sea. The Ebro Valley had just been born.

All the framing mountain chains completed their orogenic development during Oligocene times and enclosed inside their walls the Ebro Basin, which filled up gradually with considerable, though variable, thicknesses of Oligocene lacustrine deposits. With the Oligocene the orogenic history of the Ebro Valley and marginal chains drew to its end. After the end of the Oligocene there was mass lifting of the marginal mountain ranges relative to the deposits that fill up the Ebro Valley. The Pyrenees have risen as a chain of mountains, and this newly created range stimulated erosion.

It is important to focus attention on the characteristics of these two areas, the intermediate areas being much more simple. Both of them will be visited and their understanding will be made easier and more complete if their general outlines are previously known.

THE AREAS TO BE VISITED

(*i*) *Their Importance in the Search for Oil.* Areas (a), (b), (c) and (d), because of their greater thicknesses and more continuous sedimentary history, constitute distinctive zones within the general sedimentary setting of the Ebro Valley. They show lateral changes in lithology, pinchouts of strata, marginal facies developments, and a greater abundance of sandstones and they are the zones with the most hopeful prospects for oil within the region as a whole. Some of them, for example (a), contain important limestone reefs. Others, for example areas (c) and (d), contain sand and sandstone beds of very good porosity in the Albian stage. All of them show more or less abundant oil seepages, particularly areas (a) and (b). Oil prospectors will look with special interest at these areas and at their contacts with neighbouring districts. This does not mean, however, that the rest of the Ebro Valley is excluded from oil possibilities; it means only that these areas are the outstanding ones.

A relationship of cause and effect is shown in the existence of these special areas and in the concentration of the plastic and saliferous materials of the Keuper at their margins. This concentration must be attributed to the gravitational pressure due to the accumulation in them of greater thicknesses of sediments. The materials of the Keuper escaped from underneath in search of areas of lesser gravitational pressure, and migrated gradually towards the marginal zones. The manner in which the Keuper found its way to the surface, and the time, continuity or lack of continuity, of the upward movement depended on the particular characteristics of each area. In Area (a), the Keuper broke out and flowed over the surface wherever the sedimentary cover was very weak. The upward movement seems to have been reinforced by, and restricted to the period of, the Alpine Movements. In Area (b), the upward movement started when the great accumulation of Lower Cretaceous sediments was completed and persisted with variable intensity until the end of the Oligocene. The salts and clays of the Keuper ascended in a series of very deeply rooted domes through great thicknesses of sediment. The movement was much more independent of tectonics than in the first area. The domes developed haphazardly on monoclines, on synclines, or on anticlines, and, within these, either on the culminations or on the flanks. The upward movement was predominantly gravitational. Areas (c) and (d) are much less well known from these viewpoints. The salt tectonics, if any, are much more subdued.

This diapiric interplay also has its interest and importance in oil prospecting. The diapiric materials may form barriers to migration, and thus exert an influence on the formation or displacement of oil accumulations. On the other hand, the diapirs may be, in themselves, petroliferous structures, particularly the salt domes.

(*ii*) *The Montsech Area* (*a*). The sedimentary evolution along this depression is clearly shown in the cross-sections of Fig. 4. Although these sections have been derived from different sources their legends have been unified. The thicknesses cannot be exactly compared since cross-sections 1, 3, 4 and 7 are only schematic. Nevertheless, all of them have been adapted more or less to the same scale.

Section 1 shows the sedimentary situation immediately to the east and outside the depression. The series includes only a very thin representative of Upper Cretaceous lying between the Eocene and the Keuper. The Mesozoic strata are extremely schematic.

The Upper Cretaceous is present in all cross-sections, but the thickness and stratigraphic complexity increase to the west, and diminish to the south and south-west.

Section 8, similar to section 1, shows a very schematic sedimentary sequence made up of only a very thin part of the Upper Cretaceous lying between the Eocene and the Keuper.

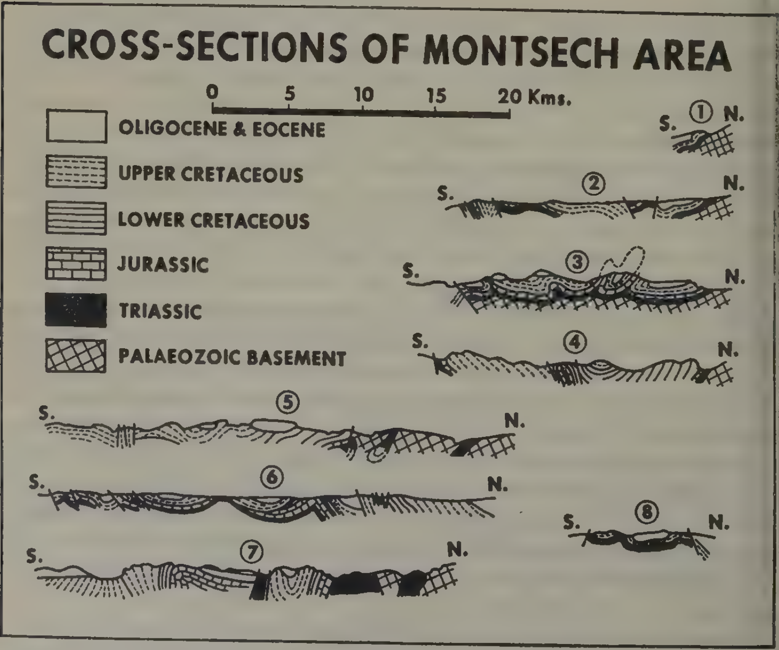

Fig. 4

The Lower Cretaceous makes its appearance between section 2, where it is still absent, and section 3. It reaches its maximum development in section 4. The thickness and complexity is already smaller in sections 5 and 6, and following the same trend as the Upper Cretaceous, the Lower Cretaceous wedges out towards the south. Section 7 corresponds to the last outcrops of the Lower Cretaceous in the west; it is completely absent in section 8.

The Lias is present, but very thin, in the zone traversed by section 2. It reaches a normal development in 4 to 7 and also wedges out to the south and west.

From all these observations we infer that a special zone of sedimentation was built up during the Lower Cretaceous (Fig. 2) as a deep depression with steep slopes giving rise to a thick accumulation of marine deposits. This basin was limited to the south by a continental mass or shelf ('Macizo del Ebro') and to the east and west by two salients ('Umbral del Ampurdan' and 'Umbral Alto Aragones'). Communication with the open seas must have been on the northern side, over the not yet existing Pyrenees. These shelves must have been created at the end of the Jurassic and erosion must have abraded the Jurassic sediments previously deposited on them.

The existence of these massifs is, so far, purely speculative and an inference from the analysis of the cross-sections and facies, but the geophysical studies and the few wells drilled to date in the Ebro Valley tend to confirm their existence.

The analysis of the character and thicknesses of the sediments which fill the depression shows a migration of the line of greatest depths towards the south from the Lower Cretaceous onwards.

The down-sinking of the massifs or shelves began gradually immediately after the Lower Cretaceous. The most durable and persistent of them all was the Ebro Massif as all the sediments of the Upper Cretaceous thin, or perhaps wedge out completely, in a southerly direction.

The greater weight of the sediments on the centre of the depression must have expelled the Keuper laterally from underneath and it accumulated towards the south where the sedimentary cover was much weaker. At the beginning of the Alpine folding, and with its help, the Keuper must have broken easily through the thin cover of sediments, favouring the folds overthrusts, and an extensive though not very intensive diapirism took place.

(*iii*) *The Cantabrian Trough* (*b*) (Fig. 4). The evolution of the sediments in this depression is clearly shown in cross-sections 1 to 5 in Fig. 5. They are all of them directly comparable, with all vertical and horizontal distances drawn to the same scale.

Cross-section 1, the most easterly, shows how the Mesozoic sequence in the region immediately to the east is made up of only a thin Upper Cre-

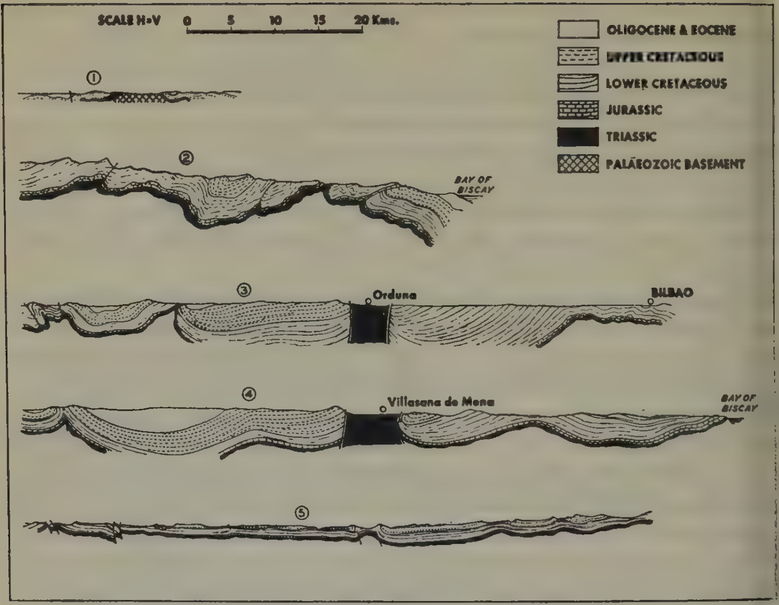

Fig. 5. Cross-sections of Cantabrian Trough

taceous representative which lies directly over the Trias and under the Eocene. Section 1 is similar (although the Upper Cretaceous sequence is more complete) to the ones on both sides of the Montsech Area (sections 1 and 8 in Fig. 4), and corresponds to the general stratigraphic sequence of the Pyrenees to the east and south of the Montsechs, and between the Montsechs and the Fosa Cantabrica.

In section 2 (Fig. 5) not only is the Upper Cretaceous much thicker than in section 1, but there is already a thick sequence of Lower Cretaceous and in addition the Jurassic (mostly Lias) makes its appearance between the Lower Cretaceous and the Trias. Both of them, Lias and Lower Cretaceous, make their appearance immediately to the west of section 1, with thin sequences at the beginning, but rapidly increasing, until the Lower Cretaceous soon exceeds 1500–2000 m.

Section 3 shows the zone of maximum development of the Lower Cretaceous.

In section 4, a reduction of thickness in the Lower Cretaceous approaching normal regional values can be observed. The Upper Cretaceous, on the contrary, reaches its maximum stratigraphic complexity, which is very great, and also its maximum thickness.

Section 5 shows a reduction of the thicknesses of both Lower and Upper Cretaceous to normal regional values.

Fig. 3 shows the position of the deep depression which corresponds to the thickest Lower Cretaceous section.

As happens all along the Pyrenees, the axis of the depression migrates from its original position during the Lower Cretaceous towards the south. We also observe the disposition of the salt-domes along a line parallel to the margin of the depression but well inside its borders. These domes, as can be seen in Sections 3 and 4 of Fig. 5, are very deeply seated.

The upward movement of the Keuper is essentially due to gravitational forces and was started, once the thick deposits of the Lower Cretaceous were laid down, at the same time as the axis of the sedimentary depression migrated towards the south, corresponding with the thickening of the Upper Cretaceous sediments in this direction.

The salt-domes originate along the line of very rapid thinning of Lower Cretaceous sediments to the south and east. This line must perforce correspond to a big fault, or rather a series of parallel faults *en echelon*, in the Palaeozoic basement underneath.

The upward movement persisted during the Eocene and Oligocene. Some of the salt-domes outcrop today in Oligocene areas; some in areas of the Upper or Lower Cretaceous. The location of the diapirs is independent of fold tectonics.

EDITOR'S NOTE. *This account was prepared and distributed to those members attending the Field Meeting. The Report of the Meeting will be published in Part 4 of this Volume.*

Summer Field Meeting in the Spanish Pyrenees

13–25 September 1961

Report by J. M. RIOS and J. M. HANCOCK

Director: J. M. RIOS

1. INTRODUCTION

WITHIN THE SPANISH PYRENEES several divisions can be recognised (Fig. 4).[1] The Axial Zone includes the highest peaks, and consists dominantly of the Palaeozoic formations of the basement. This Axial Zone is shared by France and Spain, and the frontier follows the line of the main watershed over most of its length. The southern flank is partially covered by unconformable Mesozoic formations. Due to differences in structure and lithology, these often make a distinct topographical unit—the Marginal Chains (Sierras Marginales). To the south of the Marginal Chains and along a great part of the total length of the Pyrenees, extends a narrow, elongated synclinorium, filled with folded Palaeogene sediments, which over most of the region forms lower ground—the Palaeogene Syncline or Intermediate Basin. Still farther south, the southern margin of this basin is bordered by a discontinuous chain of mountains, sharply defined and elongated east–west, which are the Subpyrenaic Mountains (Sierras Subpirenaicas). They are built of folded Mesozoic and in a lesser degree Eocene and Oligocene sediments. They never expose the Palaeozoic basement, and the lowest horizon visible is that of the Keuper. The Subpyrenaic Mountains mark the northern border of the broad Ebro Basin which contains a great thickness of Oligocene–Miocene sediments.

2. DAILY ITINERARIES

Most of the party, including the President, left Victoria on Tuesday, 12 September, by the morning boat-train, and travelled overnight from Paris to arrive at Puigcerda on Wednesday morning. There they were met by several members who had arrived the day before, and were joined by the remainder during the course of the day—one of the last to arrive being the Director himself, after a tiring coach journey which had covered the greater part of the following twelve days' route in reverse! In all there were forty members of the party including the coach-driver and the courier from the Spanish travel agency—Viajes Conde.

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[1] See *Proc. Geol. Ass., Lond.*, Part 3, p. 368.

Thursday, 14 September

Under a blazing sun, which was to remain unclouded for the greater part of the excursion, the party left Puigcerda in the coach following route N152 (towards Ribas de Freser) as far as the steep section of the road to the east and above the village of Alp. This district is in the heart of the Axial Zone. The view to the north shows a background of Hercynian granite mountains. In front of these is a downfaulted basin filled with Upper Tertiary and Quaternary on which Puigcerda is built. South of this basin, where the coach had stopped, are sediments of the basement, irregularly foliated phyllitic shales with siltstone beds, unfossiliferous but believed to be Cambro–Ordovician. There is an intricate Alpine faulting superimposed on a simple Hercynian folding.

The coach climbed the Coll de Tosas (1865 m.) and descended through Ribas de Freser on the edge of a porphyritic granite; travelling down the Freser River, the party observed from the coach the red Permo-Trias in places overthrust by the Silurian.

Where the succession is complete, the Permo-Trias is known to be overlain directly by the Garumnian[1] and all the intermediate stages of the Mesozoic are absent. Moreover, at this eastern end of the Montsech area, as illustrated in Fig. 2 of the Introduction (*Proc. Geol. Ass., Lond.*, Part 3, p. 363), there is no Intermediate Basin and one passes southwards straight from the Marginal Chains on to the Ebro Basin.

About two kilometres north of Campdevanol a stop was made near the base of the Lutetian—laminated black marls weathering white, with occasional bands of chert nodules.

From Campdevanol the route lay west along a minor road through Gombreny. The road is largely parallel to the strike of the Lutetian—blue marls, calcareous sandstones and sandstones. At the Coll de Marolla (1100 m.) the coach stopped for the party to see the view farther west. The pointed peaks of Mount Pedraforca are of marls and limestones ranging from the Aptian to the Cenomanian. Underneath the Eocene of this area, the Cretaceous has been thickening from east to west. This is the beginning of a major depositional basin extending west and south-west, and feathering out to the north, east and south.

At Guardiola the route turned south to follow the Llobregat River. From the coach could be seen the extensive mining for the hard lignites in the Garumnian. South of Serchs a stop was made to examine a roadside

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[1] The term Garumnian is used only in the Pyrenees. The type locality is the Montagne d'Ausseing in the Petites Pyrénées of Haute-Garonne, and includes both fresh-water and shallow marine sediments. It is recognised that the Lower Garumnian is upper Maestrichtian, the Middle Garumnian is Danian and the Upper Garumnian is Montian (Abrard, 1948; Casteras, 1957) and, from a recent re-study, Szöts (1959) concluded that the Lower and Middle Garumnian should be regarded as Cretaceous and the Upper Garumnian as Tertiary. In Spain it ranges from upper Maestrichtian to Cuisian (see Rat, 1959, 828) and is best regarded as a formational name for the dominantly fresh-water series (with minor marine intercalations) above the main marine Cretaceous and below the main marine Eocene.

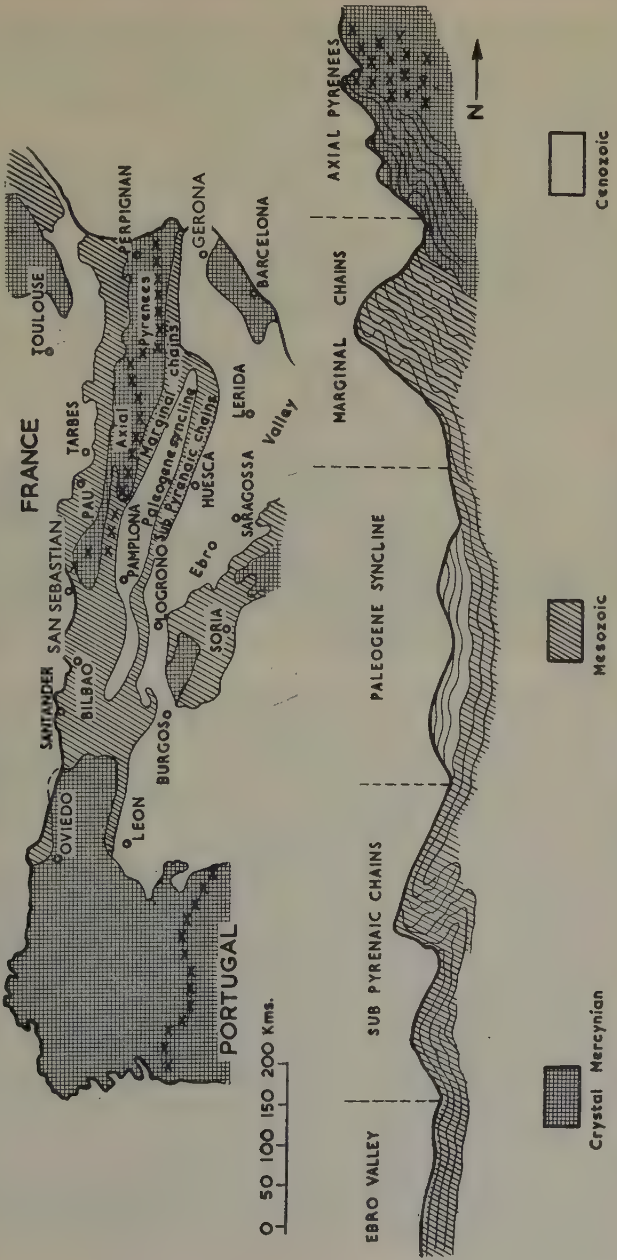


Fig. 6. General geological structure of the Pyrenees

exposure of a rich bank of rudists in the Upper Cretaceous. Hippuritids up to a foot long and three inches in diameter are mainly concentrated in one bed, although in the limestones above and below they are plentiful. They are not in the position of growth and although complete examples are rare, they are not usually crushed or badly damaged.

On the northern margin of the Ebro Valley basin lies Berga, and as the coach descended into the town, the folded marginal conglomerates of the Oligocene could be seen beside the road.

After lunch in Berga, the hair-pin minor road leading to the monastery of Queralt defeated even the skilful efforts of Señor Irazusta, the coach-driver. Moreover, the road through the Sierra de Odén was closed and so it was necessary to take a longer route via Solsona and Basella to follow C1313 beside the Rio Segre northwards to Seo de Urgel, which was no reached until about 9 p.m.

Friday, 15 September

Most of the day was spent on a southward traverse from the Axial Zone at Seo de Urgel to the northern margin of the Ebro Basin at Oliana, the road following the course of the Rio Segre (see Fig. 7 a).

Immediately west of Seo de Urgel on road C1313 a stop was made to examine massive red conglomerates, probably Pliocene, of the Seo de Urgel tectonic basin, and Lower Palaeozoic phyllitic shales of the basemen (Axial Zone).

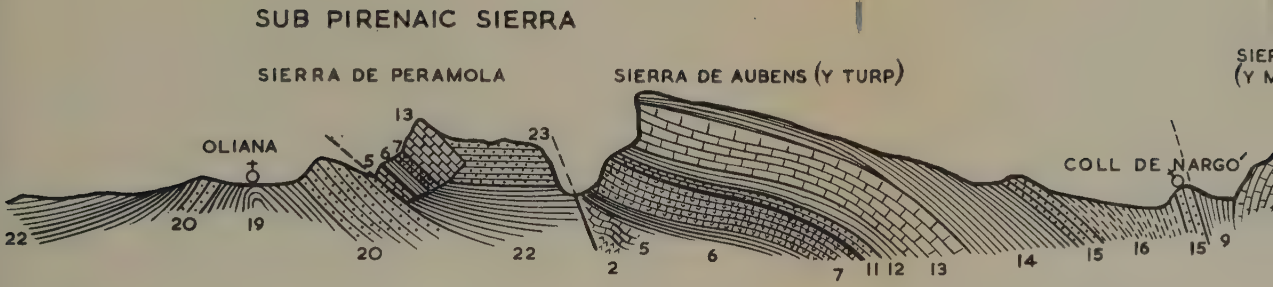
Some kilometres south the road enters red hills of continental Permian-Lower Trias, superficially resembling the New Red Sandstone of Britain. Examination showed markedly even bedding, both in massive beds and in the finely bedded ones within them. Largely of sand-grade detritus, there is a wide range from fine micaceous sands (with obscure ripple-marking) through micro-breccias to occasional pebble beds.

Farther south, the red hills are replaced by white-yellow hills of Keuper (the Muschelkalk is thin), much disturbed as so commonly seen in the Pyrenees. Where the party stopped to look, it was a chaotic deposit with gypsum tectonics—marl, limestone and other lithologies, sometimes in a breccio-conglomerate, but sometimes showing streaky foliation.

South of Tost y Hostalets members collected belemnites and brachiopods from Pliensbachian–Toarcian limestones with thin clays.

North of the westerly tributary of the Segre, a little north of Orgaña, the party examined blue marls and marly limestones of the Albian–Cenomanian in which a few Orbitolines were found.

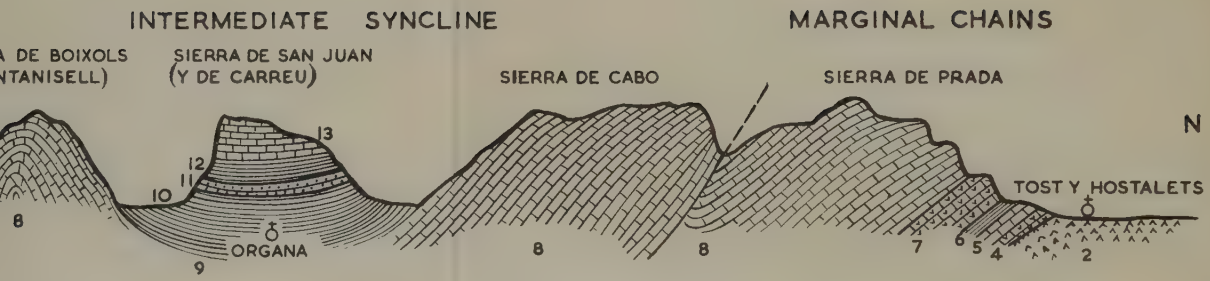
Immediately south of Orgaña a stop was made to look at the Sierra de San Juan which is capped by Upper Senonian limestones forming the core of a westerly plunging syncline, which expands farther west to form the Intermediate Basin.



1. Muschelkalk—thin-bedded limestones
2. Keuper—gypsiferous marls
3. Rhaetic—gypsum
4. Rhaetic—cargneules and dolomites
5. Lias—compact limestones
6. Lias—fossiliferous marls
7. Dogger—dolomites
8. Urgo–Aptian—limestones
9. Aptian–Albian—marls and limestones

10. Cenomanian—ma
11. Turonian—sandst
    limestones
12. Coniacian–Santor
    sandy limestone
13. Campanian—lime
14. Maestrichtian—m
15, Maestrichtian—sa
16. Garumnian—mar

Fig. 7 a. Section along the line of the Rio Segre south of Seo de U



and limestones
es and sandy

—sandstones,
nd marls
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16. Garumnian—conglomerates
17. Ypresian—clays and red marls
18. Lower Lutetian—alveoline limestones
19. Middle Lutetian–Bartonian—marls
    and flysch
20. Oligocene—basal conglomerates
21. Oligocene—basal gypsum
22. Oligocene—normal facies
23. Oligocene—marginal conglomerates

el. Length of section approximately 26 km. (*After Rios, 1956*)

[*To face p.* 376

North of Coll de Nargó the party walked along the road examining ternating marly limestones and shales of the Albian–Cenomanian, the oundaries being ill-defined, the whole often much disturbed tectonically. ossils were scarce but an *Inoceramus*? and a belemnite were found.

Close to the village a vertical fault brings the Garumnian down against e Albian–Cenomanian. The Garumnian here is a massive conglomerate light-coloured limestone pebbles in a fine red matrix. This coarse limeone detritus indicates a strong contemporary relief, presumably assoated with late Cretaceous earth movements.

Between Coll de Nargó and Oliana there was no time for any halts but e great thickness of Upper Cretaceous shales, marly limestones and limeones was obvious from the coach. On the southern flank of the Interediate Basin the Cenomanian and the whole of the Lower Cretaceous are sent. Immediately north of Oliana it was possible to see gently folded ligocene marginal conglomerates.

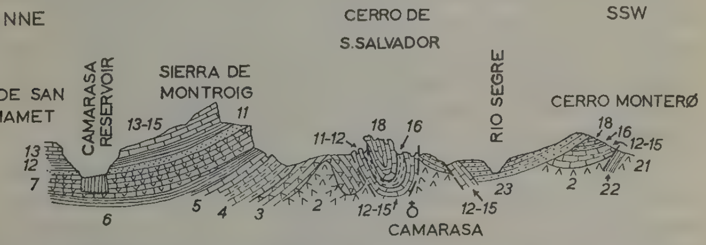
After a leisurely lunch in Oliana, the southward journey was continued er the fresh-water Oligocene of the Ebro Basin. The even, finer grained posits, above the marginal conglomerates, are not flat-bedded but gently ted and folded. The Director suggested that some of the folding may have en contemporaneous with sedimentation.

Balaguer was reached only a short while before darkness.

Saturday, 16 September

The morning was spent amongst the Subpyrenaic Mountains where they e cut by the Rio Segre north of Camarasa, twelve kilometres northrth-east of Balaguer.

Leaving Balaguer on the minor north-bound road C147 east of the River gre, the coach crossed Quaternary sediments and terraces. Some kiloetres to the north, the view to the west showed a semi-diapiric anticline white gypsum-rich Oligocene in which the original bedding is destroyed. the southern side the bedded white Oligocene is sharply upturned by the apir. Farther south, around Balaguer itself, the red Oligocene has only a ntle southerly dip.



g. 7 b. Section through the Camarasa district north of Balaguer. The Cerro de San vador is a structure in the foreground of the remainder of the section

Turning back towards the north one was confronted by cliffs of the Subpyrenaic Mountains. In the foreground is the violent syncline of the Cerro de San Salvador with Lutetian limestones in the core. This structure is probably caused by diapiric upwelling of the gypsum-rich Keuper which can be seen forming a sharp anticline immediately west of the Cerro de San Salvador. Oligocene conglomerates are banked up against the structure and form the lower flank of the hill.

In the mountains of Montroig, behind and to the north-west of the Cerro de San Salvador, the local succession is displayed: the limestone forming the summit cliffs is mainly Campanian; the Turonian–Coniacian is not prominent; the dolomites of the Dogger (Middle Jurassic) show as dark cliffs half-way up; the Lias is largely hidden by vegetation and the Rhaetic shows white near the base of the mountain.

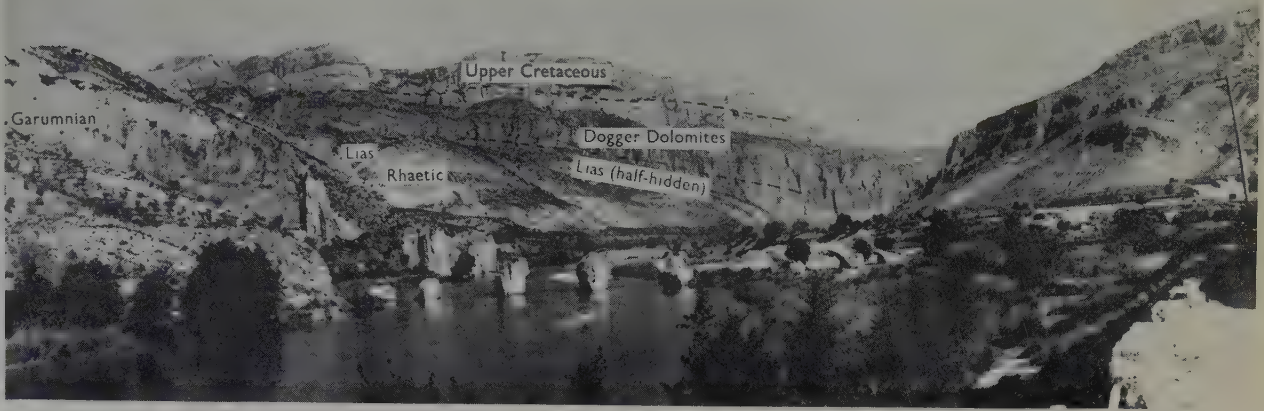
Leaving the coach on the north side of Camarasa village, the party worked down the succession towards the Camarasa reservoir. The marginal Oligocene conglomerates dip very steeply south, uptilted by the Keuper; they are well bedded and of very variable grade, the pebbles including Lutetian limestones with alveolines. The most prominent part of the sixty to eighty metres of underlying Garumnian is conglomeratic, including limestone and dolomite conglomerates. The pebbly nature of the rock is sometimes only brought out by weathering.

Beyond a break in the roadside exposures is a magnificent dolomite solution-breccia believed to be Rhaetic.

The route now became wooded and even closer to the river. Marly limestones of the Toarcian with brachiopods were exposed before long, and rather monotonous sections of red Dogger dolomites, thick-bedded and generally fine or medium grained showing early stages of solution brecciation. The coach had been driven to the reservoir and on the way back picked up members scattered along the road.

After lunch in Balaguer the party drove northwards on C147 towards Tremp. Stops were made to see the Keuper gypsum and near the convent five kilometres south of Avellanes to see ophites brought up by the Keuper. At the Coll de Ager attention was divided between soft drinks and the alveoline-rich limestones of the Lower Lutetian. Looking north from the coll can be seen the Montsech Mountains, the most northerly of the Subpyrenaic ranges in this region. The Montsech range is a huge fold, thrust southwards over the Eocene, but the lower limb is invisible along most of its length (see Figs 4 (6) and 7 c).

The party left the coach again where the road first crosses the River Noguera–Pallaresa and then follows the gorge northwards, eighteen kilometres south of Tremp. By the time the coach picked members up again they were spaced out along several kilometres of the road. The most southerly hundred metres or so are in Lutetian greywackes with numerous



N. B. Peake photo

3rd day. 1 km. north of Camarasa. View north-westwards across the Rio Segre towards the Sierra de Montroig (ruined bridge across river marks former road, now replaced by the one visible in bottom right-hand corner).

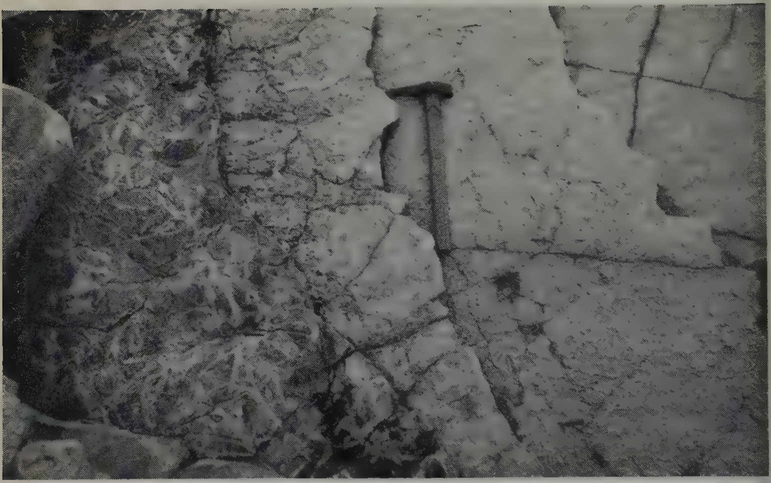
Sequence: Upper Cretaceous: massive white limestones, capping sierra. (Note that the dotted line is well above the base of these, which is partly obscured by talus.)

Dogger (= Middle Jurassic) Dolomites, characteristically reddish

Lias: Charmouthian–Toarcian fossiliferous marls

Rhaetic, which includes a great thickness of Gypsum

(The vertical beds in the left foreground are Garumnian—mainly continental deposits, including marls and limestones, representing the Maestrichtian–Danian–Montian—forming the steep limb of an anticlinal fold to the left of the present picture)



N. B. Peake photo

12th day. Limestone bands of the upper Senonian at Zumaya. Their upper surfaces (facing the camera) are covered with track- and burrow-markings, and also exhibit ripple-marks in many places. The surface to the left of the picture bears structures strikingly similar to the so-called '*Spongia paradoxica*'—well known from the Red Rock at Hunstanton, and elsewhere. (The whitish markings are actually cylindrical tubes, of lighter-coloured material than the matrix, and intertwine and cross one another—although their '3–D' structure is not very evident in the photograph) organic markings and occasional echinoids. In Lutetian limestones farther along the road there is graded bedding on a conglomeratic scale which shows the beds to be overturned. The main range itself starts abruptly with Dogger dolomites.

It was dark by the time the coach reached Tremp, set in the Intermediate Basin here known as the Conca de Tremp.

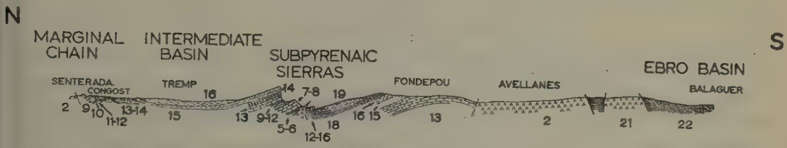


Fig. 7 c. North–south section from Senterada to Balaguer. Length of section approximately 62 km.

Sunday, 17 September

The morning was spent on the Upper Cretaceous of the Marginal Chains. Driving north from Tremp on road C147 the route followed the River Noguera–Pallaresa as far as Pobla de Segur, where the main road was left in order to continue northwards along the upper of two minor roads. Unluckily a landslip had carried part of the road down the valley side, and it was not possible to get the coach as far as was intended; members walked on as far as they could in the time available. This allowed a traverse from the lower part of the Maestrichtian to the Lower Santonian, the whole series dipping south. From the most northerly point reached (about one kilometre south of where the two minor roads and river meet) the view farther north shows horizontal Oligocene conglomerates resting on steeply dipping Upper Cretaceous.

The roadside exposures were disjointed, often covered by extremely coarse recent scree. The Lower Santonian was of grey marly limestones, very fossiliferous, and the *Inoceramus* sometimes silicified. The commonest fossil was *Micraster corbaricus* Lambert (cf. *M. brevis* Desor, of which it appears to be a southern counterpart), indicating an horizon near that of the lower third of the *Micraster coranguinum* Zone of Britain (Lower Santonian). *Inoceramus cuvieri* J. Sowerby, *Spondylus*, sponges, corals and a texanitid ammonite were also found.

The Upper Santonian, of alternating marls and hard marly limestones, was less fossiliferous but yielded a nautiloid, *Spondylus* and ? *Terebratulina*.

The Campanian was poorly exposed but there were numerous sections in the probable Maestrichtian—sandstones and clays, the sandstones fine grained, micaceous, with ripple-bedding and numerous sole-marks.

On the way back to Tremp for lunch, the party looked at the Upper Maestrichtian—massive calcareous sandstone.

Most of the afternoon was taken up with the long drive to Huesca, but several stops were made alongside the steep and winding road west of Tremp to collect large foraminifera and gastropods from the Lutetian clays and limestones. At the first of these stops, about four kilometres west of Tremp, the commonest species was *Turritella trempina* Carez and less abundantly *T. rodensis* Carez.

Monday, 18 September

From Huesca the coach drove north-west along road N240. Near Ayerbe a stop was made to photograph extraordinary pillar-erosion in red Oligocene conglomerates. A little farther north the road cuts through the Sub-pyrenaic Mountains. This area is in a salient north of the Aragon massif which must have risen above sea-level during the Upper Jurassic, allowing erosion of earlier formations and impeding sedimentation until it was re-submerged during the Senonian. Consequently the succession is much simpler than that in the Montsech region to the east, and the Senonian rests directly on the Keuper. Part of the Oligocene is upturned and overthrust to the south, but much of it is post-orogenic and almost flat lying.

A stop was made at the Pantano de la Peña at the north end of the gorge where the Lower Eocene grey-blue cavernous crystalline limestones rest on Garumnian red marls and limestones. Some beds of the Eocene are rich in alveolines and nummulitids.

Continuing northwards the route led across the Intermediate Basin as far as Puente la Reina where the coach turned east through Jaca and beyond on road C134. Before the Gallego River is reached the road turns north into the Marginal Chains. Owing to the combination of limited thickness (about 800 m.), incomplete succession (Coniacian on Carboniferous) and a steep southerly dip, the outcrop of the Mesozoic is quite narrow. At Biescas one is still on Tertiary; south of Polituara one is already on Devonian.

The newly constructed road climbing up to the hotels at Panticosa is not only steep and narrow but includes several extreme hair-pin bends. After lunch the party was able to examine the surrounding Panticosa granite, and again beside the road on the way down. It is known that there is a wide range of facies within the granite (see Van Landewijk, 1960), but immediately north of the hotels it is largely a medium-grained leucocratic granite. Loose boulders sometimes show poorly developed biotite clots suggesting an almost complete assimilation, but the majority of the numerous xenoliths suggest that the granite was both cool and poor in volatiles at the time of intrusion. The xenoliths are angular, show no reaction between granite and sediment, no decrease in crystal size in granite at contact with xenolith, no larger crystals in granite under large xenoliths, very little metamorphism of sediments even at the contact. Calcareous

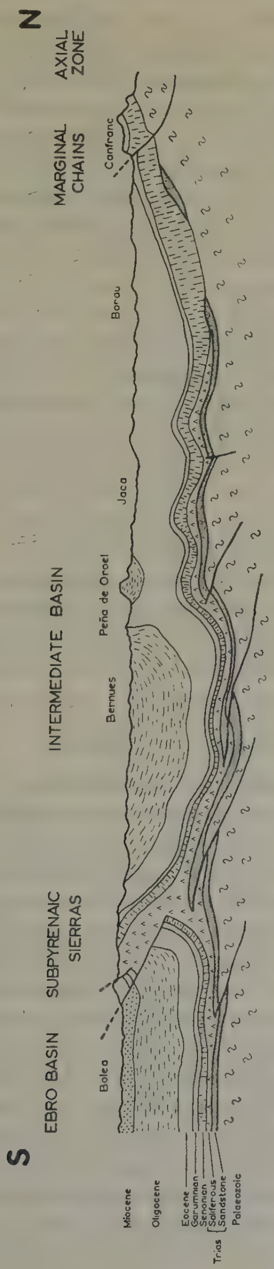


Fig. 8. North–south section from Canfranc to Bolea, north of Huesca. Length of section about 55 km. (*After Selzer, 1934*)

rocks are barely recrystallised to marbles. Yet near the margins of the granite, looking at the east wall of the river gorge from the approach road, one can see folding within a limestone series pass undisturbed into the granite (care should be taken in interpreting such views of limestones; much apparent vertical bedding is only vertical markings made by water washing down over the surface of the rock). Mapping of the region has shown a ghost stratigraphy in the granite. The folding of the limestones in this gorge also illustrates a general tectonic pattern of the area—a cascade of knee-folds descending from the divide on the Franco–Spanish frontier to the foot of the Marginal Chains.

At the fortress of St. Elena the party examined the Danian white algal limestone in the quarries across the bridge to the east of the main road. About half a kilometre farther south the coach was stopped again to allow members to look at some Lower Eocene alternating shales and graded clastic limestones with well-developed fracture cleavage in the shales at some 60° to the normal bedding.

It was dark by the time the party got back to Jaca.

Tuesday, 19 September

There was no arranged programme, members being left free to go sightseeing or to do independent field work. Some spent most of their time in the hotel, this being the one day with bad weather; the rain was continuous after noon, and in the morning light of the following day was seen to have fallen as snow on the higher sierras!

Wednesday, 20 September

The morning was spent in the valley of the head-waters of the River Aragon north of Hecho, amongst the Marginal Chains. At the village of Siresa the party looked over the ancient church with its windows of alabaster. There were also good exposures of the Eocene flysch.

A little farther north the Mesozoic forms imposing cliffs in which the tectonic style is perfectly displayed; a pile of flat-lying folds with approximately east–west strike stand out by the contrast between the white Danian limestone and the darker Maestrichtian marls and sandstones.

Farther north still there were a number of points where one could see the unconformable contact between the Coniacian and the red Permo-Trias. This was particularly clear in the view north-west from the Oza forest.

Until recently the people of the village of Hecho, where the party lunched, wore their traditional colourful costume. Examples of the clothing are still preserved in the local town hall; after lunch two members of the party were dressed in this old style, to the entertainment of the rest, who were also able to look round one of the old houses in which all the original fittings and furniture have been preserved.

At the main road, N240, the route lay west towards Pamplona. On one bend a car coming fast in the opposite direction took the corner too closely, and its driver, seeing the coach in front of him, overcompensated his steering, shot off the edge of the road and overturned. Members of the party helped to free the occupants and turn the car the right way up again. Just before the tributary of the Esca the turning to the north through Sigüés was taken. North of the gorge through the Sierra de Leyre, and beyond the village, the party examined the Lutetian whilst the coach was driven on farther through the Sierra de Navascués to turn round. The Lutetian here is of alternations of calcareous sandstones a few inches thick and siltstone beds one to two feet thick, the shaly siltstone beds showing the early stages of fracture cleavage. The sandstones show features which some members considered characteristic of flysch—there was much discussion in the party on the use and meaning of the word 'flysch'—unsorted graded detritus which includes felspar grains and a variety of clastics.

On the way back a stop was made at the northern end of the gorge. This is in the eastern end of the Sierra de Leyre, a Subpyrenaic Chain consisting of an anticlinorium thrust southwards over the blue 'Canal Marls' and variously faulted. The Maestrichtian was seen to be a fine siliceous sandstone overlain by dark limestone with calcite fragments. The overlying Garumnian is continental—red clays with green patches and occasional beds of white limestone. This thin sequence is capped by a pyritous mudstone followed by thick massively bedded limestone.

Back on the main road the westward drive to Pamplona was resumed.

Thursday, 21 September

The morning start was delayed whilst an unsuccessful attempt was made to trace the driver of a car which had struck the rear of the coach during the night. The route then followed was back towards the east, but this time by minor roads to the town of Aoiz on road C127. A little north of this town and overlooking to the north the lowland of the confluence of the Urrobi and Irati rivers, is a roadside cliff in Eocene. The sequence is rhythmic; each limestone bed has an upper part in which there is current bedding shown by the sand content and is often immaturely slumped (convolute bedding). But the impressive feature here is large-scale slumping producing not only large folds but shattering beds of limestone to form a slump conglomerate with blocks a metre or more across.

The drive continued northward through the Urrobi gorge in the Marginal Chains. One to two kilometres south of Nagore a roadside traverse was made beside the Maestrichtian and Danian. The Maestrichtian, dipping north, is reddish, sandy, slightly current bedded, coarsely shelly massive limestone. The overlying Danian, exposed north of a bridge over a small eastern tributary, is a white to purple-grey limestone.

South of Arce, near kilometre post 38, is a succession of southward dipping Senonian marly limestones with a rich fauna of echinoids, accompanied by a lesser number of ammonites and lamellibranchs, particularly *Inoceramus*. Members collected keenly over the slopes of the hill and the finds included: many specimens of *Micraster* ex gr. *aturicus* Hébert (a southern counterpart of *M. brongniarti* Hébert) including *M. corcolumbarium* Desor and *M. gourdoni* Cotteau; several examples of *M.* (*Isomicraster*) aff. *faasi* Rouchadzé, as well as some early *Cyclaster*—perhaps *munieri* Seunes. *Echinocorys* was well represented by very large specimens whose shapes (cf. *turrita* Lambert, *carinata* Desor, *héberti* Seunes, *tenuituberculata* Leymerie, and *subglobosa* Goldfüss) are all closely confined to the top of the Zone of *Gonioteuthis quadrata*, and perhaps the base of the succeeding zone, in Britain or Belgium. *Inoceramus goldfussianus* d'Orbigny (= *I. balticus* Böhm), characteristic of the same horizon, was likewise common, and specimens of *Spondylus* and numerous sponges were also collected. The seven ammonites collected are of less value than the echinoids, the preservation allowing only the following determinations: *Bostrychoceras* cf. *polyplocum* (Roemer); *Neancyloceras* sp.; *Pachydiscus* s.s. sp. nov. ? aff. *llarenai* Wiedmann; several indeterminable pachydiscids. Wiedmann's holotype came from the highest Maestrichtian of Zumaya, but this assemblage is more likely to be high Campanian.

The fossils were collected from a hundred or more metres of sediment which possibly embrace several zones.

Continuing northwards the coach-driver at first missed his westward turning through Espinal and had to turn in a meadow which was softer than he had expected. It was necessary for the party to provide a stone foundation to the wheels before the coach could be driven back on to the road, and thence along the picturesque road C135 to Pamplona for lunch.

In the afternoon a visit was made to a shallow quarry about half a kilometre off the road near Cizur, some six kilometres west of Pamplona. This is in the topmost Lower Bartonian thinly bedded shelly limestone rich in polyzoa. The President, Mr. Pitt, comments: 'The polyzoa are too poorly preserved to be specifically determinable, but it is noticeable that the ratio of cyclostomes to cheilostomes is higher than one would expect in a Tertiary sediment.'

The rest of the afternoon was taken up with the drive to Vitoria, although on the way it was possible to see one of the derricks for oil-drilling in this district.

Friday, 22 September

Most of the day was spent in the Sobrón gorge cut through Lower Cretaceous to Oligocene, with Upper Cretaceous particularly well developed and exposed (see Fig. 9). To reach the gorge the coach drove south-west from Vitoria along the N1, and turned north-west at Miranda de Ebro just before the Ebro River; the river was now followed upstream so that on road N625 the coach took the westward turning towards Sobrón, running along the north side of the river.

At the first sections on the north side of the road the party left the coach, not rejoining it until the middle of the afternoon. The gorge is cut through an anticline with a north–south strike and whose eastern flank has dips approaching the vertical. The youngest beds examined were Oligocene conglomerates, sandstones and mudstones, pink, well-bedded detritus in a typical molasse facies.

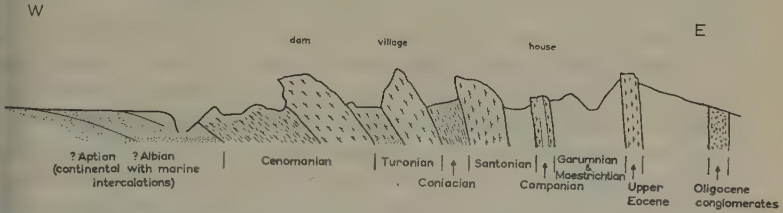


Fig. 9. Eastern half of the Sobrón Gorge, 35 km. west-south-west of Vitoria. Length of section from the axis of the anticline to the Oligocene conglomerates is 3.5 km.

The exposures of the Eocene are mainly limestones, the softer lithologies being covered with vegetation. One of these limestones forms the most easterly of the great ribs of rock which dominate the scenery of this valley. The depression immediately west of it is formed by the Garumnian—yellowish calcareous sandstones and dolomites.

In the roadside below a large house is a rubbly limestone rich in fossils, probably high in the Campanian, possibly Lower Maestrichtian. Members collected orbitoids, rudists and other lamellibranchs.

The underlying Santonian is more thickly developed and most of the stage is well exposed. Topped by evenly bedded, fine grained dense calcareous sandstones, it passes down through a series of sandstones, sandy limestones and marls, some levels with echinoids, lamellibranchs and *Lacazina*, the base being formed by a great rib of limestone just east of the recently built hotel.

The Coniacian is of soft marly limestone forming a depression opposite the hotel. There is an oyster bank in the middle of the stage and other lamellibranchs occur sporadically.

Opposite the older part of the village, where some of the party ate their packed lunch, the sparry limestone of the Turonian forms another great rib.

The Cenomanian limestones begin in the depression to the west of the Turonian rib and east of the new dam. The limestone here is deeply discoloured by weathering. The outcrop of the Cenomanian is comparatively wide, not only because of a greater thickness, but because the dip is flattening as one approaches the axis of the anticline.

It is not until one gets beyond a large road bridge that the Albian is reached. This is in a completely different facies from the rest of the Cretaceous—massively bedded felspathic sandstones with strongly developed current bedding within each massive bed. Some beds show good examples of sorted graded bedding. Immediately west of the bridge the grade ranges from silts to pebble beds, but as one moves farther west, i.e. down the succession, the grade becomes finer, eventually to dark shales. Probably most of this pre-Cenomanian series is fresh-water, but in one bed, probably Aptian, specimens of an echinoid (?*Hemiaster*) were found.

This was as far as most of the party got, members being spaced out all along the road when the coach picked them up again to drive to Salinas de Añana, about four kilometres east of road N625. This village is famous for its production of salt obtained by channelling a saline stream into a series of sun-heated evaporating pans, said to date from the Roman era onwards, constructed along the valley sides. The stream is fed by a spring emanating from a salt-plug, which, starting in the Keuper, has punched through some five miles of sediment in the centre of a large syncline. The surface levels have long been leached of salt but the structures resulting from the original diapiric movement are still well displayed. The Oligocene of the district is gently folded but at the margin of the plug it is sharply upturned. On the east side of the village the party saw basic igneous masses brought up by the salt, and about one kilometre east of the village there is an old quarry displaying perfectly salt flow tectonics, with a variety of xenoliths in a matrix streaked around them.

The party returned to Vitoria for the night.

Saturday, 23 September

The coach started south-westwards on the N1, turned on to a minor westerly road some ten kilometres from Vitoria and then north at the village of Subijana. This attempt was shortly defeated by a tunnel with a bend too narrow for the coach which had to be backed to Subijana, whence the westward road through Salinas de Añana was taken. At the N625 the coach headed north and drove over a plateau of limestones and marls—mainly Coniacian–Santonian—to the Orduña pass.

The route followed on this day is approximately illustrated by section 3 of Fig. 5 (*Proc. Geol. Ass., Lond.*, Part 3, p. 370). The same section in more detail is given in Rios, 1956, fig. 22. From these sections, and also from section 4 of Fig. 5, it can be seen that the zone of maximum sedimentation migrated southwards during the Cretaceous, so that the maximum thickness of the Lower Cretaceous is north of Orduña and that of the Upper Cretaceous well to the south of it.

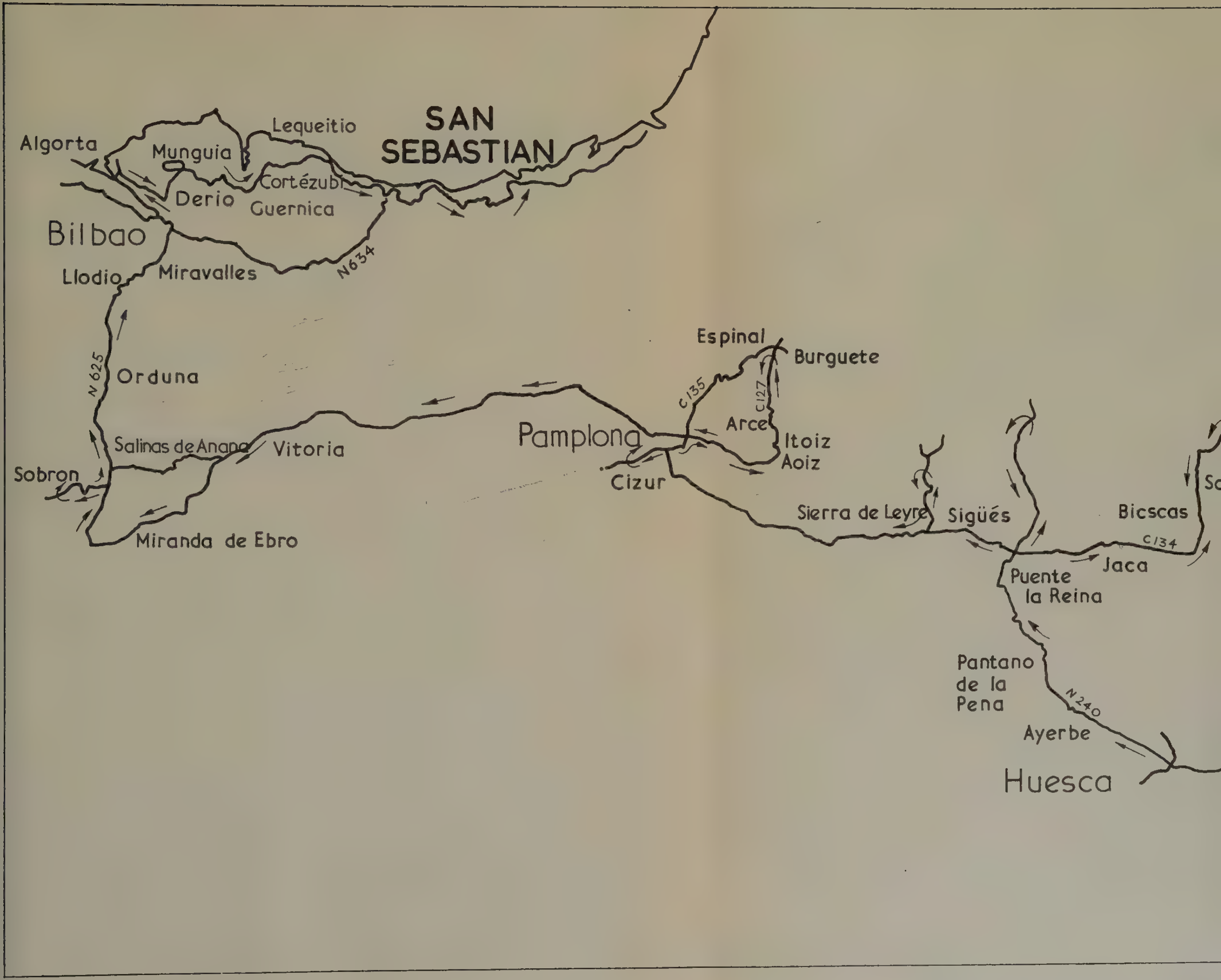
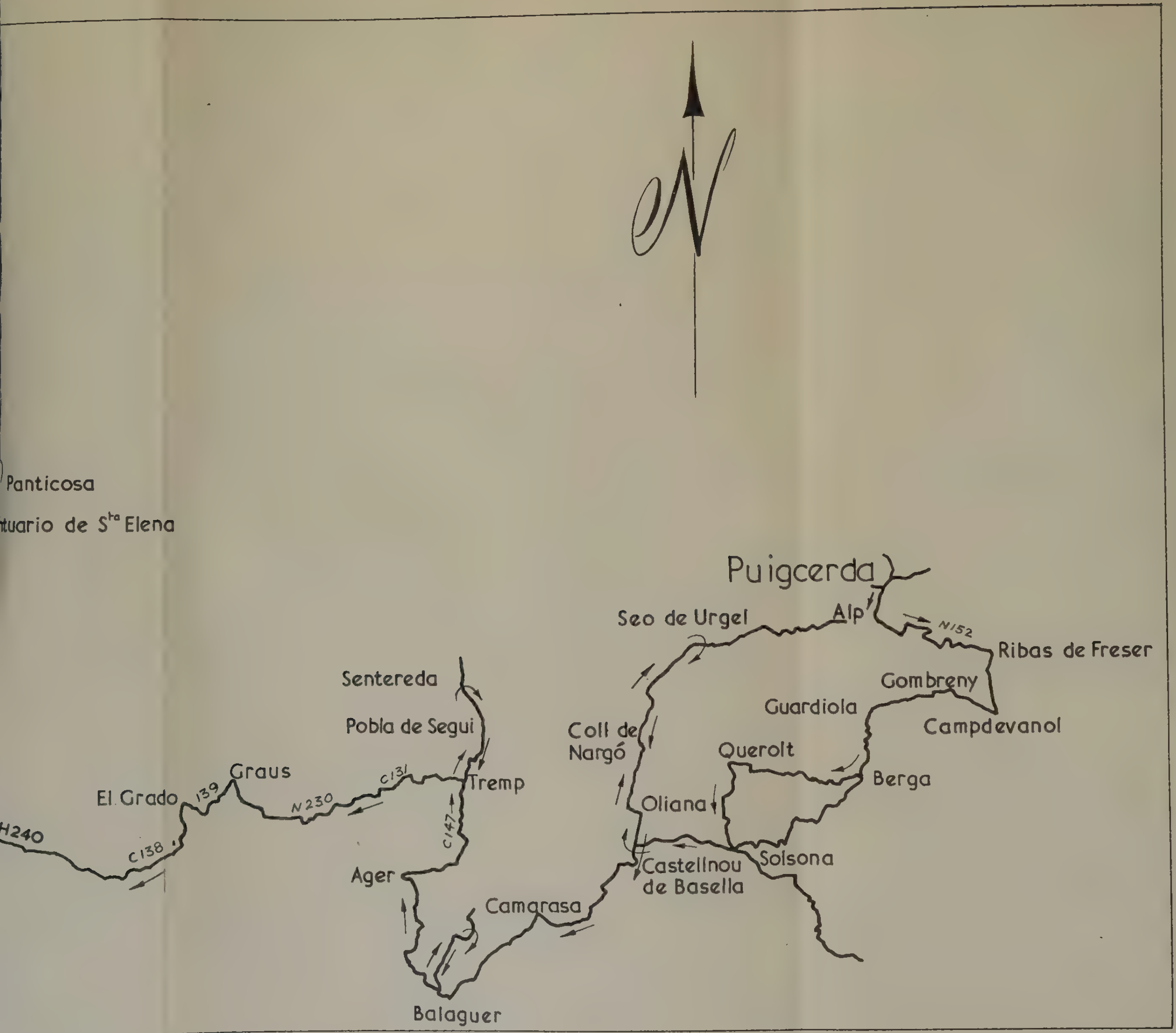


Fig. 10. Map showing the route of the pa

All sections are diagrammatic and to no



y and principal localities mentioned in the text

rticular scale.

[*To face p.* 386

Orduña itself is placed within an enormous circular flat-bottomed de-
ession largely surrounded by steep or even vertical walls of Cretaceous
estones. The depression is of Keuper brought up several miles by a
t-plug. A short way down the winding road the party left the coach,
st to admire the view and consider the general structure; then to walk
rt of the way down the hill looking at the Upper Turonian—very thick
onotonous silty grey limestones with poorly developed bedding, and
parently no fossils other than the occasional burrow.

After lunch in Orduña the northward drive was resumed. At Amurrio
road bank exposure showed the Cenomanian to be well-bedded
spathic, richly micaceous calcareous sandstones. Farther north near
yando a lower level in the Cenomanian is in a facies looking like shelly
ndovery of the Welsh Borderland—massively bedded grey sandstone
athering yellow-brown with a lamellibranch–gastropod fauna.

In a large quarry near Llodio is exposed black to blue-black Aptian or
bian limestones. This limestone is marginal to an actual reef facies, but
n here the beds are about ten feet thick, and within each massive bed
re is no subsidiary bedding. At several levels rudist reefs are developed
t owing to the solid massive nature of the rock the type and position of
rudists are not clear. On weathered surfaces the rudists are preserved in
ck limestone which contrasts with the light grey matrix. The dip is steep
the south-east and the thickness even within this quarry must be
asured in scores of metres.

Just south of Miravalles roadside exposures of Lower Cretaceous flysch
re examined. Micaceous silty sandstones with oscillation ripple bedding
interbedded with different thicknesses of sandstones. Graded bedding
are, and no sole-marks were seen.

As the coach was entering the suburb of Bilbao it was halted by a large
ry across the road. A group of men was failing to push it back until
ne of the members of the party lent their weight.

Sunday, 24 September

The drive at the start of the morning zig-zagged from Bilbao to Las
enas and Algorta by the estuary road, thence back south-east to the
ss-roads near Derio, north-east through Derio along C6313 to Munguia,
th-east again to Guernica, then the direct road towards Lequeitio.

Near Cortézubi a stop was made to look at a rudist reef, probably
bian in age. The unbedded black limestone contained many examples of
rudist *Horiopleura baylei* (Coquand), both complete and broken, and
als apparently in the position of growth. The matrix contained much
ll-pebble grade reef debris. Cavities in the rock are sometimes rich in
erals including calcite, bitumen and ? fluorite.

Lunch was taken in the open air at Lequeitio and then the eastward drive was continued to a point six kilometres west of Zumaya. Here the roadside cliff exposes a thick succession of well-bedded Upper Cretaceous fine-sandy and silty limestones and marls, but no fossils other than tracks were seen. At the western outskirts of Zumaya another roadside cliff exposes steeply dipping Maestrichtian? flysch with graded bedding which shows it to be inverted. The micaceous sandstone–shale succession shows an enormous range of organic sole-marks, but *Inoceramus* is also common and a possible scaphitid was seen.

High tide prevented the party from examining the Danian rose limestone on the coast at Zumaya, but on the eastern outskirts of the town an old quarry showed Danian red marls alternating with light grey limestones. No fossils were seen.

Some kilometres east of Zumaya a stop was made to look at the magnificent exposures beside the road of Eocene flysch. This is basically a series of alternating sandstones and micaceous silts. The sandstone beds are the thicker, ranging up to about a metre, whilst the siltstones are thinner, not exceeding thirty centimetres and generally much thinner. Each bed of sandstone or silt is well bedded within itself but graded bedding was not seen. Each sandstone bed starts with a selection of sole-marks; flute casts are particularly common and a variety of trace fossils, including: *Lorenzinia apenninica* De Gabelli, *Münsteria bicornis* Heer, *Scolicia prisca* Quatrefages, *Subphyllocorda* (= *Palaeobullia*) sp. and a type without a Latin name but corresponding to the 'Rhabdoglyphen' of Fuchs. The main part of the bed is finely laminated by the profusion of muscovite. Some bedding planes are rich in carbonaceous plant fragments. The top of many of the sandstone beds shows convolute bedding for a thickness of several cm. The junction with the flaggy silts is quite sharp.

In the evening at San Sebastian a formal dinner was held. The President expressed the thanks of the whole party to Dr. Rios for all the work he had done to make the meeting such a success and presented him with a silver tray which was to be suitably inscribed. Presentations were also made to the courier Señor Vargas and the coach-driver Señor Irazusta.

Monday, 25 September

There was no set programme for this day, and indeed some of the members who were travelling home independently left early in the morning. The rest enjoyed the shops and beach in San Sebastian, examined the Eocene north of the city or journeyed back to Zumaya to look at the shore sections of the Upper Cretaceous.

The cliff section at the foot of Mount Urgull, north of the Aquarium, provides a magnificent exposure of Eocene flysch. The succession here is mainly in sand grade but includes pure limestones with sharply truncated tops, indeed, there are sometimes pebbles of limestone in the overlying sandstone. There is very little graded bedding, and that in the thinner sandstone beds within silty flag intercalations. The main sandstone beds are well laminated, the upper part of the beds often showing large-scale convolution ripple bedding or immature slumping such that there is no cut-out of the laminations as in true ripple bedding, although this is also common in some beds.

In the evening the coach returned to take the party to the frontier at Irún and across to Hendaye station, whence the train journey home began.

ACKNOWLEDGMENTS

The authors are grateful to several members of the party who have helped with this report. Dr. L. R. Cox has identified lamellibranchs, gastropods and trace fossils; Mr. N. B. Peake has provided identifications and comments on echinoids; and the President, Mr. L. Pitt, has commented on the polyzoa. In addition Mr. N. B. Peake and Dr. H. Reading have read the manuscript and made a number of useful criticisms. We are also grateful to the Editor, Mr. G. T. Raine, for preparing the locality map, Fig. 10, from information provided by Mr. Pitt.

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Differentiation in the Olivine-Gabbro of the Insch Mass, Aberdeenshire

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ABSTRACT: Geological, petrographic and geochemical accounts of two of the basic units of the Insch Igneous Mass are provided. One, the Insch Differentiated Suite, consists of iron-enriched olivine-gabbro with up to 24% total iron-oxides, that passes upwards through syeno-gabbro into syenite, with over 7% potash. This suite is interpreted as resulting from the seeping of low-temperature constituents towards the roof of the crystallising olivine-gabbro magma which, itself, may have been enriched in iron by the abstraction of refractory magnesium-rich phases from the regional gabbro magma. The second basic unit dealt with is the Bourtie Series of dunites, peridotites and troctolitic gabbros, all rich in magnesia, and interpreted as early cumulates. The possibility of the geological connection of the two units to make a wider differentiated series ranging from dunite to syenite is discussed. Whilst this interpretation is favoured, it is agreed that special geochemical and mineralogical investigations are necessary to test its validity. Thirty-seven new analyses are used in the petrogenetic discussions. A revised geological map of the whole Insch Igneous Mass is presented.

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1. INTRODUCTION

THE INSCH Igneous Mass is the largest of the Newer Basic intrusions of north-east Scotland that were emplaced after the regional metamorphism of the Dalradian country-rocks and before the deposition of the local Middle Devonian. The exposed area of the Mass is some eighty-five square miles, of which eighty square miles is occupied by gabbros and their derivatives. Of these basic rocks three units can be distinguished:

(1) A series consisting of olivine-gabbro, orthoclase-gabbro or syeno-gabbro, and syenite that can be demonstrated to result from strong gravity differentiation *in situ* of gabbro magma. This series we may call the 'Insch Differentiated Suite'.

(2) A group of dunites, peridotites and troctolites making the eastern part of the Insch Mass in the parish of Bourtie. It is possible that this

'Bourtie dunite-troctolite Series' is a differentiation cumulate from the Insch olivine-gabbro and that the two rock series so far mentioned together make one grand differentiated series ranging from dunite to syenite.

(3) A group of 'hypersthene-gabbros and norites' that in our opinion finds no place in any straight differentiation-series from an olivine-gabbro parent. These noritic rocks are of considerable interest and their origin will be discussed in a later communication.

In addition to the dominant gabbros, the Insch complex includes in its western part a small development of granitic and dioritic rocks that have been described by Sadashivaiah (1954a). The dioritic rocks are considered to result from the contamination of a granitic magma with basic material. This granite is quite unlike the syenitic derivative of the gabbro magma and was clearly not directly produced from it. Most likely it forms a thin sheet between the basic rocks and their roof.

The general distribution of the Insch igneous rocks is shown in Fig. 1. This is the first modern map of the Insch Mass as a whole, the mass being divided on existing Geological Survey maps between Sheet 86 (1923) and Sheet 76 (1886). The Insch ground, apart from certain localised areas, is not well exposed and the construction of a solid geological map is consequently a matter of some difficulty. Much of the boundaries between the igneous components of the mass is based on soil-types, topographic features and other indirect evidence. Residual gabbro boulders, as Blundell & Read (1958) have shown, can be utilised for mapping purposes by those experienced in field-work in north-east Scotland. There is, fortunately, a multitude of small quarries and many of these provide material of petrogenetic significance. But, all the same, the internal boundaries shown on the map cannot be claimed to be anything more than approximate. Further, the connection between the Insch Differentiated Suite and the Bourtie dunite-troctolite implied in the map is, as already admitted, a matter for discussion and for further work. Nevertheless, we are satisfied that the demonstration of the passage from olivine-gabbro through syeno-gabbro to syenite that characterises the Insch Differentiated Suite is of high validity. This opinion, resulting from the field-work and laboratory study of Read and Sadashivaiah, is reinforced by the pattern of the geochemical variations provided by the chemical analyses executed by Haq.

None of the earlier investigations dealt with differentiation in the Insch gabbros. In the short Survey Memoir on Sheet 76 (1890), Hinxman grouped the Bourtie dunites with the serpentines bordering the mass on its south-west contact (see Fig. 1). But Read (1956) has now shown that these serpentines, whatever their ancestry, are mechanically emplaced along a fracture zone and are not, in their present position, related to the Insch igneous rocks. Hinxman gave also a generalised distribution of the syenite

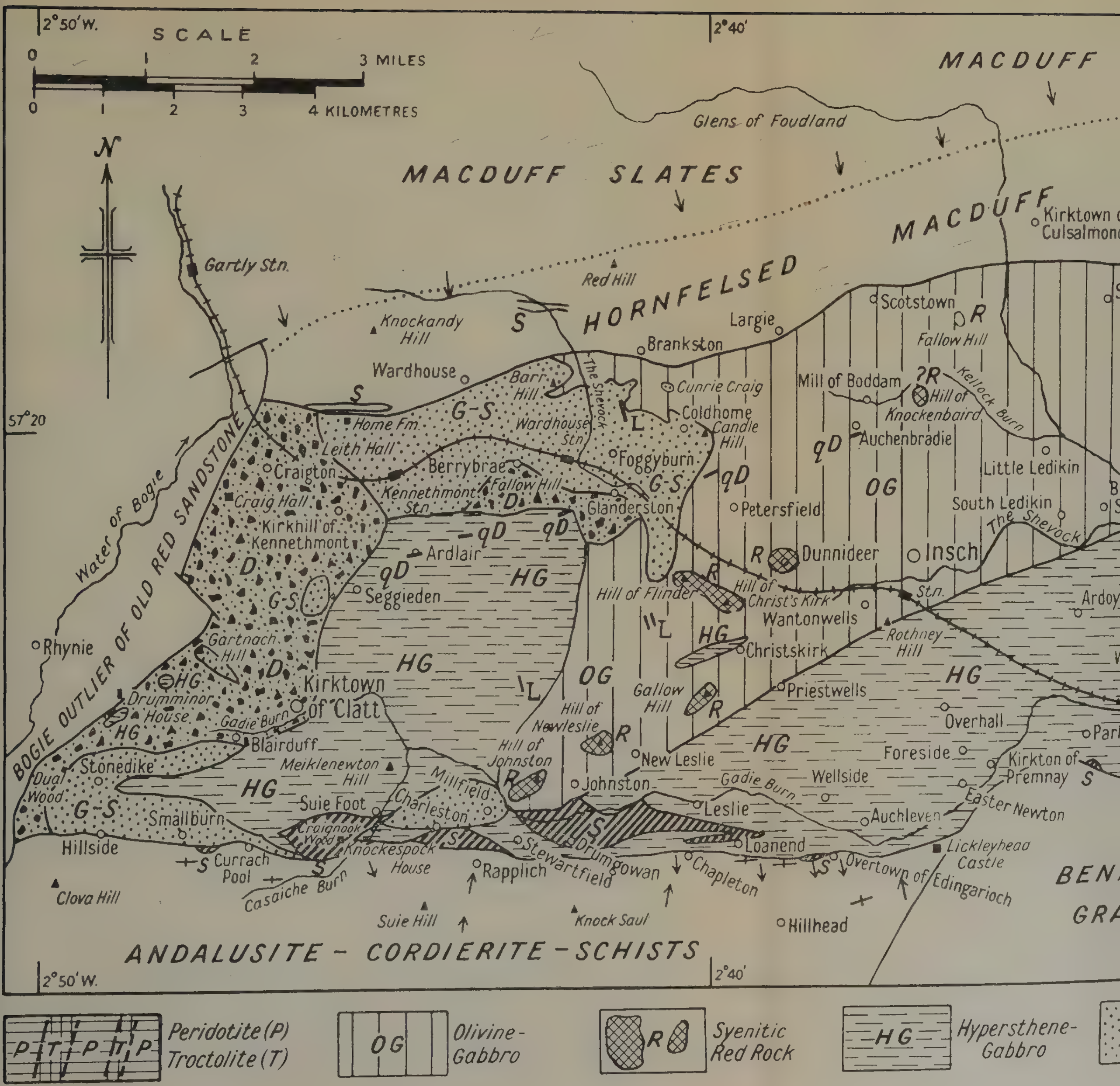
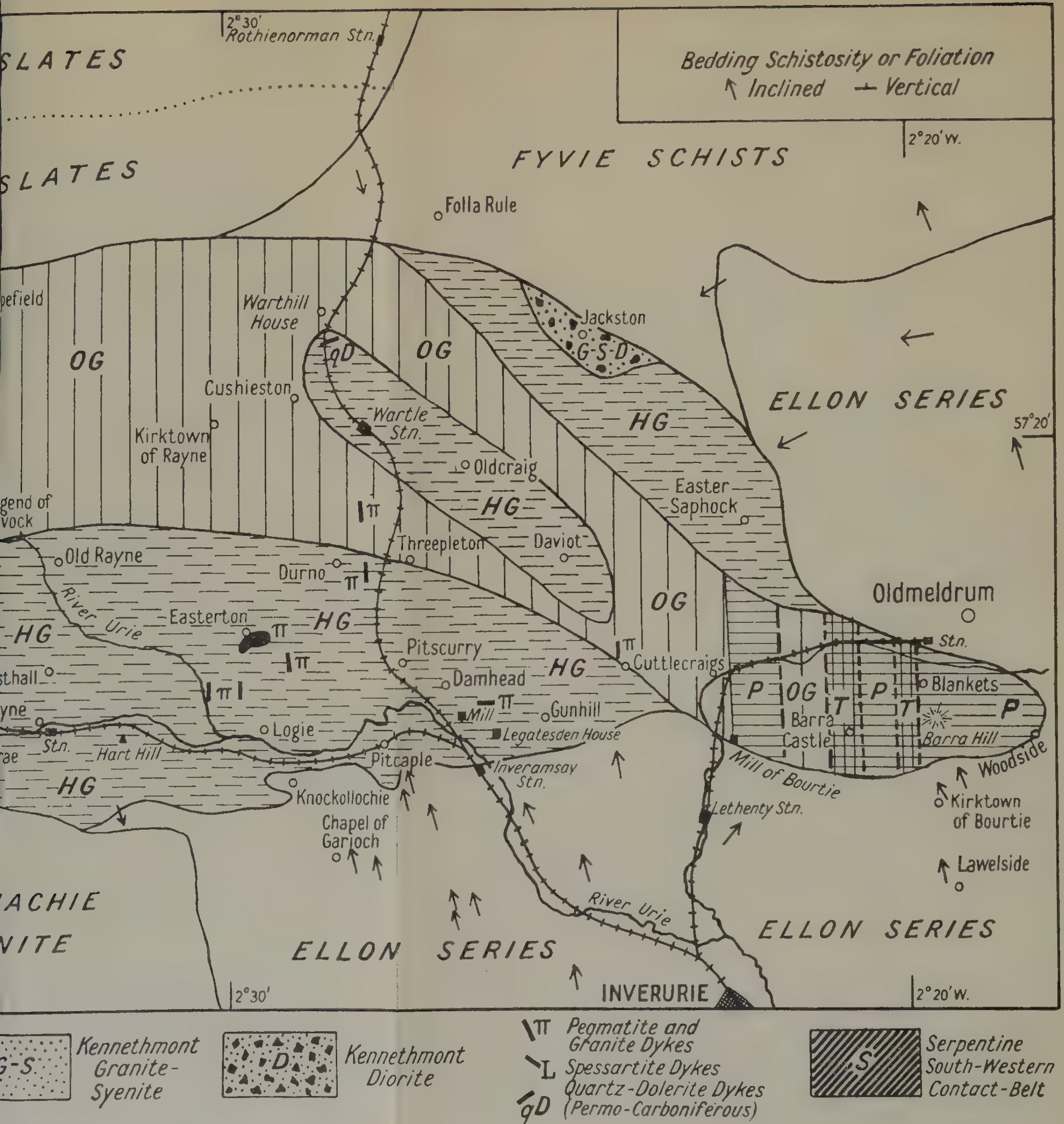


Fig. 1. General map of



he Insch Igneous Mass

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nember of the 'differentiated suite' and stated that syenite veined the local diorite' (that is, gabbro). The olivine-gabbros of the differentiated suite alling within Sheet 86 of the 1-inch Geological Survey Map were described y Read (1923) but no other members of the suite came into his ground.

Two more modern contributions are concerned with the Bourtie dunite-roctolite Series. On the 1-inch Geological Survey Map, Sheet 76, the eridotites were shown as a narrow band running east and west in the abbros, parallel to the southern contact of the Insch Mass. In 1936, G. Vhittle gave the first description of the peridotites and troctolites and uggested that the ultrabasic masses were lenticular accumulations of erromagnesian constituents on the floor of the igneous body. He agreed in eneral with the Survey outlines for the peridotites. Sadashivaiah (1954b), fter detailed consideration of all the outcrops and of the attitude of the anding in the gabbros, suggested quite a different form from that proosed by the Survey and by Whittle. He concluded that the large- and mall-scale banding ran north and south and that, consequently, the outhern contact of the Insch Mass could not be the floor on which the ltrabasic rocks accumulated. All the natural contacts of the mass were onsidered by Sadashivaiah and by Read (1951, 1956) to be roof. Although he ultrabasic rocks possibly lay nearer the bottom and termination of the nass than any of the gabbros to the west, the floor was nowhere exposed.

The first purpose of this present paper is to describe the geological ccurrence and distribution of the members of the Insch Differentiated uite and to interpret this as the result of the differentiation in place of a abbroic magma into intermediate and acid types culminating in a syenitic red rock'. Petrographic descriptions adequate for this purpose are given nd the abundant chemical data concerning the rocks themselves are used n the discussion of the broad petrogenesis of this unit. Secondly, the ourtie dunite-troctolite series is dealt with in the same fashion. Finally, he problem of the relation of these Bourtie rocks to the differentiated uite is taken up and lines of enquiry for its solution are suggested.

In the chemical work by Haq, rapid methods of silicate analysis as nodified by Mercy have been used for the major elements. The samples vere prepared for analysis by the routine procedures for avoiding conamination. The accuracy of his work was tested by Haq in making four nalyses of each of two peridotites. The estimated accuracy is similar to hat developed by Mercy (1956; see also 1959).

The distribution of the Dalradian country-rocks of the Insch Mass is hown in Fig. 1. Since these rocks are not directly concerned in the proluction of the differentiated suite, it is sufficient to state that they consist f the greywackes and pebbly grits of the Macduff Group, the andalusite-ordierite-schists and pebbly grits of the Fyvie Group and the granulites, chists and gneisses of the Ellon Series (Read, 1923, 1952; Whittle, 1936).

Excellent hornfelses are produced from these country-rocks adjacent to the Insch Mass. These are undeformed along the northern and eastern contacts but, as Read (1951, 1956) has shown, are progressively deformed westwards along the southern contact until, west of the Bennachie Granite, the whole contact zone is eliminated. Two conspicuous shear-belts, parallel to the mechanical southern contact, and minor fracture planes traverse the rocks of the differentiated suite; the modifications of the primary rocks of this suite by shearing are readily determined (Read, 1956, 83–4), and its presence in no way affects the conclusions concerning the course of the differentiation.

2. THE INSCH DIFFERENTIATED SUITE: OLIVINE-GABBRO TO SYENITE

Olivine-gabbro forms in a general way the northern half of the Insch gabbro body, extending from east to west for a dozen miles from Daviot through Rayne to Insch and Leslie parishes (Fig. 1). To the west of the town of Insch, a low but conspicuous range of hills—Dunnideer, Hill of Flinder, Hill of Christ's Kirk, Gallow Hill, Hill of Newleslie and Hill of Johnston—rises from the gabbro plateau (Fig. 2). On each of these hills there is visible an upward transition from gabbro to syenitic 'red rock' and it will be convenient to call the range the 'Red Rock Hills'.

(a) The Olivine-Gabbro

By far the dominant areal component of the differentiated suite is the olivine-gabbro. This rock was described by Read (1923, 117–19) under the title of olivine-norite; both orthopyroxene and clinopyroxene are common, the orthorhombic type usually being the more abundant. Read (1935) has attempted to restrict the term norite to gabbros in which the pyroxene is entirely orthorhombic, a restriction very convenient for the petrogenesis of this north-east Scottish gabbro province. The rocks now considered are strictly olivine-hypersthene-gabbros.

(*i*) *Field Occurrence*. The general distribution of the olivine-gabbro is shown in Fig. 1. It makes the northern contact to the east of the Shevock and, in the Hill of Johnston, approaches very closely to the present southern boundary of the Mass. In the west, olivine-gabbro builds the lower slopes of the Red Rock Hills and passes upwards, as we show later, into the syeno-gabbroic types making the upper parts of the Hills. Its relations to the hypersthene-gabbro dominant in the southern part of the Insch Mass are not considered here.

The olivine-gabbros are generally coarse dark rocks weathering into spheroidal balls, the centres of which supply excellently fresh specimens. The rock is not banded nor, except in restricted zones already described by Read (1956), is it greatly affected by the late cataclastic deformation.

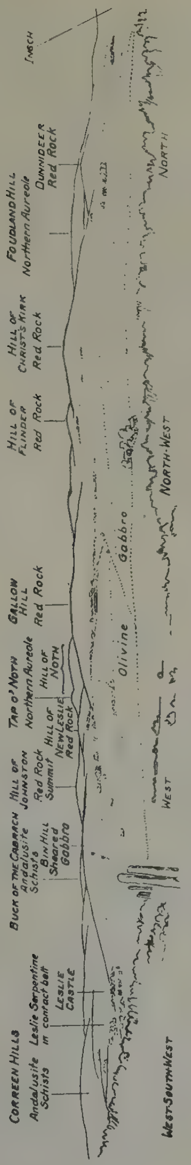
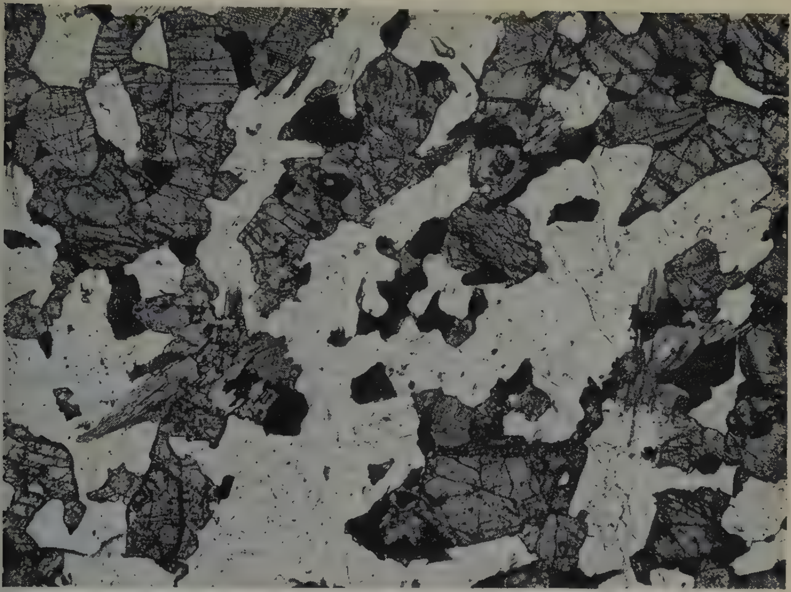


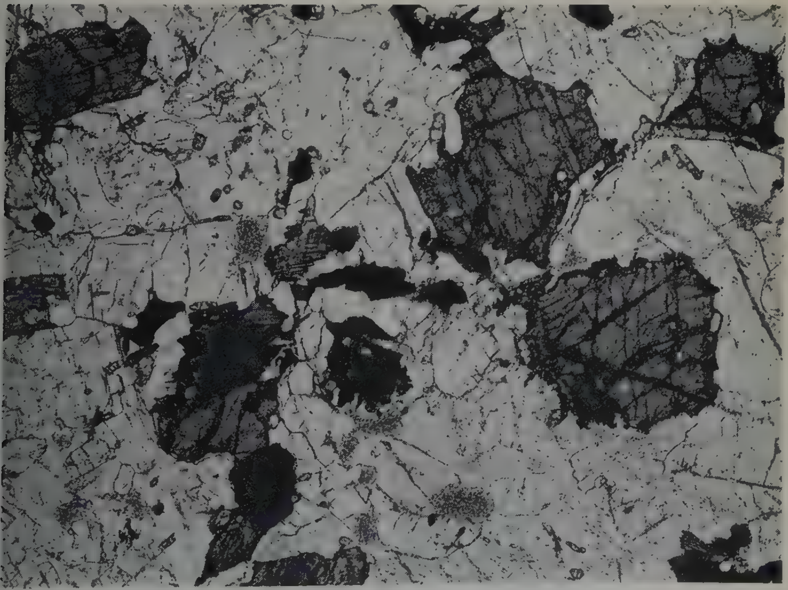
Fig. 2. The Red Rock Hills, viewed from near Wellside, two and a half miles SSW. of Insch

(*ii*) *Petrography*. Plagioclase, olivine, hypersthene, augite and iron-oxide are the main components, with apatite an abundant accessory and hornblende and biotite in small amounts (Plate 17 A). A general description of the olivine-gabbro from the central part of its outcrop around Bonnyton, three miles north-east of Insch, and Kirktown of Rayne is the following. The plagioclase is typically a labradorite, An_{57}, in unzoned crystals, small, well formed and even lath-like when enclosed in ophitic mafic plates, but subhedral and mutually interfering when larger. Olivine is the most abundant mafic mineral, occurring as grains moulded on feldspar, crudely ophitic plates or residual grains in reaction associations. In one specimen, the olivine was determined as Fo_{23}. Hypersthene is more abundant than the clinopyroxene, forming stout prisms or ophitic plates with typical hypersthene pleochroism in pale shades. It is often of Bushveld type, with lamellar intergrowths of clinopyroxene. It may enclose drops of clinopyroxene to produce, in some examples, almost a micropegmatitic texture—similar textures are considered by Wager & Deer (1939) to be the result of simultaneous precipitation of the two pyroxenes and not of exsolution. Hypersthene is a constant member of the reaction association, as described below. The clinopyroxene is dusty-green in colour, schillered, and with finely developed sahlitic structure. It forms prisms and plates with ophitic relations to feldspar, and takes part in the reaction-association. Black iron-oxide is abundant in large irregular grains; it may be bordered by hypersthene or form vermicular intergrowths with this mineral. Apatite, in large stout prisms, is a constant accessory, and at times becomes almost an essential constituent. Hornblende is scarce and in some rocks it is absent; it usually forms narrow deep-brown rims to pyroxene or olivine, especially when near iron-oxide. Biotite, pale yellow to deep reddish-brown, is associated with iron-oxide in irregular flakes and patches; occasionally it is fairly abundant as the external member of reaction-associations.

No coronas similar to those to be mentioned from Bourtie are found in the main olivine-gabbro, but intergrowths of various kinds and reaction-associations are beautifully developed. Magnetite and hypersthene form vermicular intergrowths and together replace olivine in reaction-fashion (e.g. Read, 1923, 118). The discontinuous reaction series involved in the reaction association consists of the following members: olivine, hypersthene, augite, hornblende and biotite. It is shown by the occurrence of irregular cores of olivine followed outwards by fairly continuous zones of the successive minerals of the series. Often, however, some members of the series may be missing. Iron-oxide occurs abundantly in the hornblende and biotite zones. Another texture commonly met with in the olivine-gabbro is the patchy texture, in which the plagioclase and single individual ferromagnesian minerals are associated in ophitic patches, the plagioclase crystals being smaller in size than the normal plagioclase occurring by



A. *Olivine-gabbro*, Bonnyton, Insch; plagioclase, olivine, hypersthene, clinopyroxene (sahlitic), magnetite



B. *Syeno-gabbro*, Hill of Johnston, Insch; olivine in irregular grains, scarcer pyroxene and magnetite, large plates of orthoclase enclosing euhedral plagioclase

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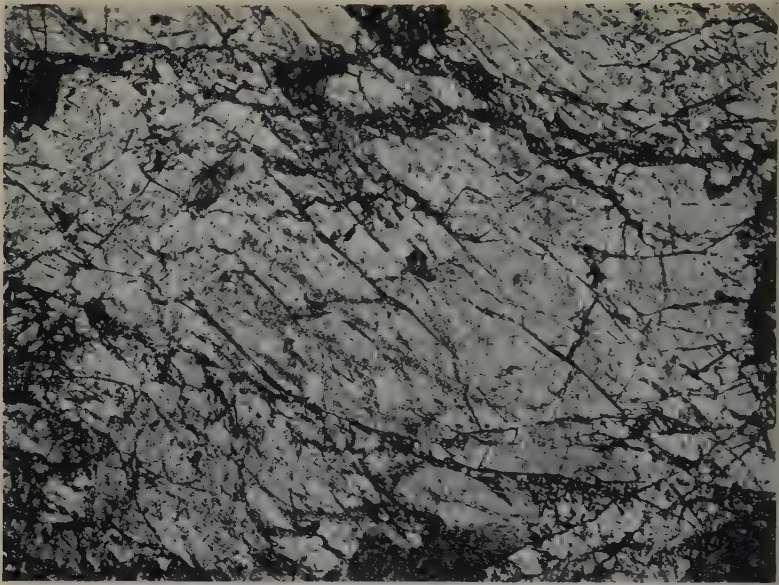
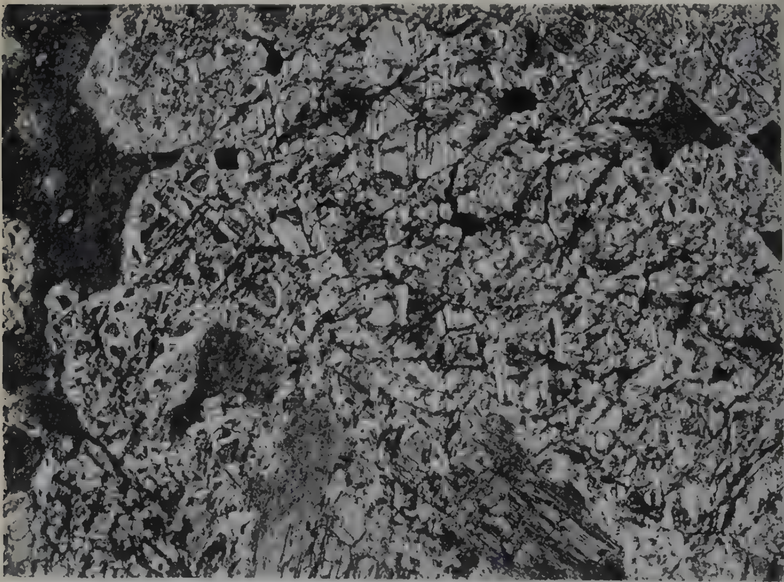


A. *Syenite* (*Red Rock*), Gallow Hill, Insch; orthoclase and oligoclase, with scanty chloritised hornblende; some brecciation



B. *Dunite*, Barra Hill, Bourtie; serpentinised olivine grains packed together, with a few interstitial altered pyroxene patches

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itself. Reaction and patchy textures exactly similar to those of the Insch olivine-gabbro have been described by Read (1935, 601) from the olivine-gabbro of the Haddo House Mass and have been interpreted as due to a relatively rapid cooling of the magma (cf. Bowen, 1928, 56, 58, 59).

Variations in the olivine-gabbro arise chiefly by changes in the relative amounts of the two pyroxenes. Occasionally, as near Threepleton and Durno, pyroxenes are almost absent, an olivine-plagioclase rock resulting which, however, bears little resemblance to the Bourtie troctolitic gabbros. Another variant is provided by the increase of hornblende and biotite in large plates enclosing residual grains of pyroxene and exemplified by rocks near Brankston in the west of the olivine-gabbro area. This latter variation is bound-up with a general east–west variation that requires detailed mineralogical and chemical investigation of the mineral phases for its proper demonstration. Such a demonstration would have a bearing on the possible relation of the Bourtie dunite-troctolite series to the olivine-gabbro (see later, p. 411). It can be said at present that in general the gabbros show, towards the Red Rock Hills, more abundant plagioclase and this of a more albitic composition. Near the Red Rock Hills, two plagioclases measured by Haq gave An_{38-40}. Further, as we have just noted, hornblende and biotite increase in amount as do, apparently, magnetite, ilmenite and apatite. Eventually, potash feldspar appears in the gabbro assemblage, the resulting rocks being considered with the syeno-gabbros dealt with in the next section of this account.

(*iii*) *Chemical Composition.* In Table I are given six analyses of olivine-gabbros made by Haq, together with that of the olivine-gabbro from Bonnyton, extracted from the Survey Memoir (*in* Read, 1923, 124). The localities of all analysed rocks are given in the Appendix. Attention may be directed to the high iron-content of the Insch olivine-gabbros as a whole, as is brought out by their ratios of Fe_2O_3+FeO to $Fe_2O_3+FeO+MgO$ given in Table I and by their positions in the $FeO+Fe_2O_3 : MgO : Na_2O+K_2O$ triangular diagram of Fig. 3. In this figure it will be seen that they form a compact cluster located towards the iron-oxide apex. This richness in iron is reflected in the high fayalite content of the olivines, in their abundance of iron-oxides and, presumably, by the ferriferous nature of their mafic minerals in general. The analyses given in Table I are arranged roughly in a geographical order of the rocks analysed from east to west, that is towards the Red Rock Hills. Thus the rock analysed as 13 is farthest away from the Hills, and those given as 18 and 19 are on the lower slopes of the Hills themselves. It will be noted that these two last rocks show the remarkable iron-oxide content of 24% and have a specific gravity of 3.2. Further, the iron-oxide content of the analysed rocks shows a general increase towards the Hills. These are matters that enter into the petrogenetic discussion of the differentiated suite that we present later (page 406).

TABLE I. **Olivine-Gabbros**

	13	14	15	16	17	18	19
SiO_2	45.6	45.1	44.65	45.3	45.5	40.2	43.0
TiO_2	2.5	2.7	3.69	3.4	3.1	3.2	3.0
Al_2O_3	16.7	16.3	17.95	17.9	18.5	12.2	9.8
Fe_2O_3	1.2	2.9	1.09	2.9	2.8	—	0.4
FeO	14.5	12.2	15.02	11.7	13.6	23.6	24.1
MnO	0.24	0.23	0.36	0.19	0.23	0.44	0.61
MgO	5.4	4.7	3.34	5.1	3.9	4.9	5.2
CaO	8.6	9.5	9.50	8.5	7.5	9.3	6.9
Na_2O	3.1	3.3	2.79	3.2	3.1	3.4	2.7
K_2O	0.46	0.59	0.54	0.50	0.60	0.60	1.9
P_2O_5	0.82	1.10	0.09	0.29	0.13	0.50	1.3
H_2O	0.51	0.58	0.55	0.62	0.73	1.30	0.55
Total	99.63	99.20	100.10	99.60	99.69	99.64	99.46
S.G.	3.04	3.18	—	3.00	3.06	3.21	3.22
$\frac{FeO+Fe_2O_3 \times 100}{FeO+Fe_2O_3+MgO}$	74.5	76.3	82.9	74.2	80.9	82.7	82.5

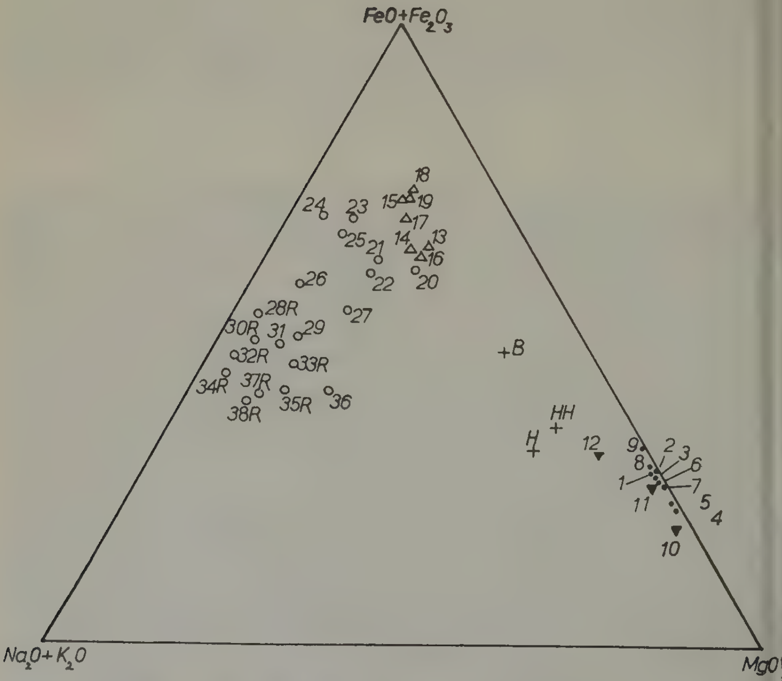

Fig. 3. $FeO+Fe_2O_3:MgO:Na_2O+K_2O$ Diagram, Solid circles, dunite and peridotite; solid triangles, cumulate gabbros; open triangles, olivine-gabbros; open circles, syeno-gabbros. R syenitic red rocks; B Belhelvie olivine-gabbro, H Huntly olivine-gabbro, HH Haddo House olivine-gabbro

(b) The Syeno-Gabbro and Syenite of the Red Rock Hills

In his topographic study of the Insch area, Glentworth (1944) delimits a 'Central Area' as a hilly tract of strong relief, with a range in altitude from 400 feet to 1000 feet, contrasting with the more gentle though high plateau to the west and the lower plain region to the east. This 'Central Area' coincides broadly with what we have called the Red Rock Hills (Fig. 2), and is the topographic expression of its geological structure, an upward sheet-like succession from olivine-gabbro to syenitic 'Red Rock'. The lower slopes of the hills are good arable or pasture, their summits are permanent thin turf, again a reflection of their geological constitution. Some of their summits reach the elevation of the northern contact, which is taken as a natural one. A view, either from the east near Insch or the west near Petersfield, of Dunnideer, conspicuous because of the ruined Garioch Castle crowning its summit, reveals the typical aspect of the Red Rock Hills; from a fairly level and extensive platform underlain by olivine-gabbro rise abruptly the residuals of the higher members of the differentiated series.

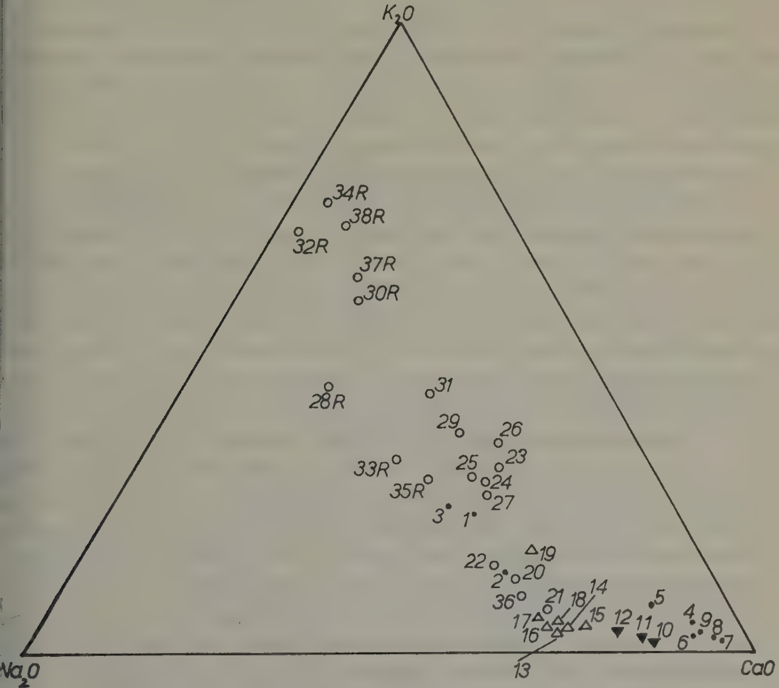

Fig. 4. CaO:Na$_2$O:K$_2$O Diagram. Explanation as in Fig. 3

(*i*) *Field Relations*. The Red Rock Hills run in a line from Fallow Hill, two miles north of Insch, south-westwards for six miles to Hill of Johnston, near the southern boundary of the Mass. The hills in order are Fallow Hill, Hill of Knockenbaird, Dunnideer, the high ground formed by Hill of Christ's Kirk and Hill of Flinder, Gallow Hill, Hill of Newleslie and Hill of Johnston. Three field-divisions can be made in the rocks forming these hills: (1) all the lower slopes and bases are of *olivine-gabbro* similar to those just described; (2) above this comes an intermediate zone of *transitional* character made of a rock of a pale brownish colour, coarse in texture and weathering like the olivine-gabbro; it is composed of tabular plates of pale brown feldspar, biotite, pyroxene, amphibole and scarce olivine and its thickness is of the order of fifty feet; we have called it syeno-gabbro or orthoclase-gabbro; (3) the tops of the hills are made of a syenitic rock, pale pinkish grey in colour, composed dominantly of pink feldspar, streaks and patches of chloritic and amphibolic material and containing abundant miarolitic cavities; it is stained red with hematite in many places and is often broken and brecciated. This acid rock is called 'Red Rock' because of its colour and of its close similarity to the famous Red Rock of Duluth (Grout, 1918); the maximum thickness of the Red Rock zone is about 200 feet. What covered it is not seen but from a variety of reasons it is thought to be near the original roof of the olivine-gabbro mass. Since the relationships between olivine-gabbro, transitional rocks and Red Rocks are fundamental in the interpretation of the differentiated series, a summary of these field-relations seen in each hill is now given.

Fallow Hill rises from an undulating plain underlain by olivine-gabbro; it is cultivated almost to the summit where there is a large quarry in a coarse-grained brownish-grey rock similar to rocks found in the other hills just below the Red Rock zone—no Red Rock itself remains on Fallow Hill, if it were ever there. Across the Kellock Burn to the south-south-west the **Hill of Knockenbaird** affords a number of poor exposures, best in a grassed bank which runs up the north face of the hill from the level of the Kellock; olivine-gabbro forms the base of the hill and is succeeded upwards by more feldspathic rocks until the summit, made of a pink syenitic rock, much crushed and quartz-veined, is reached. Two miles to the south-west, the conspicuous hill of **Dunnideer** rises abruptly from the valley of the Shevock and the general lowland of olivine-gabbro; west of Dunnideer typical olivine-gabbro is well exposed near Petersfield and, at the base of the southern slope alongside the railway, sheared and broken olivine-gabbro cut by sheared pegmatite is seen in an old quarry at a level of about 400 feet; higher up the southern slope, at a level of about 600 feet, is a small quarry in rather pale 'basic' rock, containing orthoclase, and similar rocks are found still higher up on the eastern slope; the top of the hill and the high western slopes are formed of rough-weathering, friable, coarse-grained, brownish and reddish syenitic rocks; a transition from olivine-gabbro to Red Rock can be reasonably pieced together on Dunnideer. On the other side of the Shevock Valley is the hill-mass formed by the **Hill of Flinder** and **Hill of Christ's Kirk**; owing to the extensive brecciation along the Shevock Valley to the north of these hills (Read, 1956) and also to the south in the vicinity of the farm of Christskirk, it is difficult to identify the gabbros that make the bases of the hill-mass, these basic rocks being now in the condition of brecciated and sheared greenstones; above them there appear more feldspathic and more acid types, seen, for example, north of Christskirk on the southern slopes of Hill of Christ's Kirk and on the north face of Hill of Flinder above the main road; finally, on the highest parts of the hills, brecciated and drusy Red Rock is well exposed, especially on the Hill of Flinder. The next Red Rock Hill to the south is **Gallow Hill** which provides some of the best exposures of the differentiated series; thus, on the south-western slope, practically continuous exposures extend from the base of the hill in the large roadside quarries up to the summit; at the roadside are seen feldspathic gabbros which pass upwards into the syenitic Red Rock of the higher levels; the Red Rock is much silicified, sheared and veined by quartz. West-south-west of Gallow Hill rises another Red Rock Hill, that of **Newleslie**, where the olivine-gabbro, exposed on the eastern flanks, passes upwards into orthoclase-gabbro and finally Red Rock. The last hill to be noted is the **Hill of Johnston** which provides fairly good exposures at its western end, in the large quarry and above it, and at the eastern end; at this latter locality, olivine-gabbro is found fairly high up on the slope and is succeeded upwards by orthòclase-bearing types and finally by Red Rock at the summit; at the western end, orthoclase-gabbros are exposed in a large quarry and above this on the western slopes rapid transitions to Red Rock occur.

Two final points concerning the field-relations remain to be noted. First, although it is clear that the Red Rock is a type transitional from the more basic varieties, it is often observed to vein the orthoclase-bearing gabbros and to occur in them in patches with diffused contacts. Second, so far as exposures permit an opinion, it appears that the proposed sheet structure is inclined to the west—olivine-gabbros occurring higher up on the east than on the west slopes of the Red Rock Hills.

(*ii*) *Petrography*. It is clear from the account of the field-relations just given that the best material for petrographic study is provided by the southern hills, Gallow Hill, Hill of Newleslie and Hill of Johnston. Accordingly, most attention is given to the transitional series as developed in these hills. It is proposed to give a general picture of the series, beginning with rocks that differ only slightly from the main olivine-gabbro.

The earliest stage in the differentiation is marked by the appearance of antiperthitic patches in the otherwise normal olivine-gabbro. An example of this stage is provided by a rock from the southern slope of Hill of Newleslie. The constituents of such rocks are tabular andesine-labradorite, irregular olivine grains, subhedral to ophitic hypersthene, platy pale green sahlitic clinopyroxene, together with brick-red biotite, green amphibole and iron-oxide in small amount. In a few of the plagioclase crystals there are found small irregular pools of orthoclase.

The next stage in the transition can be exemplified by rocks from the south-eastern slopes of Hill of Johnston. Olivine is abundant as shapeless or sub-ophitic grains, pyroxenes form large ophitic plates, biotite is plentiful especially in association with iron-oxide grains. Medium plagioclase is the dominant feldspar; microperthitic orthoclase either forms large irregular pools, enclosing many small plagioclase euhedra (Plate 17 B) and olivine grains, or acts as a kind of cement to the larger plagioclase crystals. Noteworthy are the accessory components; apatite prisms are much more abundant than in the normal olivine-gabbro and zircon appears in slender rods. Reaction associations of the mafic minerals are absent or developed only in the crudest fashion. A modal analysis of a rock coming from the bottom of the quarry at the south-western base of Hill of Johnston and fairly typical of this stage gave the following: plagioclase 46%, orthoclase 15%, pyroxene 15%, hornblende 8%, olivine 7%, hypersthene 3% with biotite, iron-oxide and apatite forming the remainder.

Rocks retaining a general gabbroic aspect but seen in thin section to be highly feldspathic, with orthoclase dominant over an intermediate plagioclase, provide the next term in the series. Such rocks come from the roadside quarry at the south-western base and the outcrops above it on the south-western slopes of Gallow Hill and from the Hill of Johnston quarry. The Gallow Hill rocks are seen in thin section to be made up of some 80% feldspar, together with olivine and pyroxenes, with variable amounts of biotite and hornblende, and conspicuous iron-oxide, apatite and zircon. Of the feldspars, a microperthitic orthoclase is dominant in broad plates often enclosing small euhedral plagioclase crystals of andesine composition. The plagioclase is in euhedra or subhedra and occasionally shows a poor zoning. Olivine forms shapeless, very irregular grains and may reach 2.5% of the rock; sahlitic clinopyroxene and pale-coloured orthopyroxene occur in similar grains or irregular plates; in amount they may attain nearly 4% of the rock. A deep brownish-red to pale yellow biotite varies in amount in different rocks, occasionally becoming the chief mafic constituent; it, with scarce amphibole, is sited near iron-oxide clots. Apatite is an abundant and conspicuous accessory, and zircon is noteworthy. Iron-oxide forms common irregular grains associated with biotite or olivine. No sign of reaction association is seen in the mafic components. Rarely a few specks of quartz can be detected in some rocks. A mode of a rock from the Gallow Hill quarry gave: perthitic orthoclase 47.4, andesine 35.1, clinopyroxene 2.6, orthopyroxene 1.2, olivine 2.6, iron-oxide, biotite and chlorite 10.5, quartz 0.27, apatite 1.3%. The rocks from the upper parts of the Hill of Johnston quarry are similar to those just described; a mode of one rock gave: perthitic orthoclase 30%, plagioclase 28%, clinopyroxene 16%, olivine 5%, hornblende 6%, quartz 1%, with hypersthene, iron-oxide, zircon and apatite forming the remainder. Some rocks from this same quarry contain a considerable amount of apparently primary hornblende in large plates enclosing pyroxene and olivine as irregular grains. The amphibole is deeply coloured with strong absorption and is pleochroic in yellow-brown and deep sepia.

In the next stage of the differentiated series olivine is not present and the rocks have lost the gabbroic colour and texture. Rocks of this stage are coarse in grain and pale brownish-grey in colour; they weather in rectangular blocks and are seen to be composed dominantly of large plates of pale brown feldspar with small patches of mafic constituents, pyroxenes amphiboles and iron-oxides. A rock typical of this stage comes from the west slope of Hill of Newleslie; it is composed of orthoclase, oligoclase, pyroxene, biotite, iron-oxide and zircon. The microperthitic orthoclase in large plates is dominant over the oligoclase which forms tabular prismatic crystals, sometimes enclosed in the potash feldspar. The dusky pyroxenes, mostly monoclinic, form scattered prismatic grains and are by far the dominant mafic component, biotite and hornblende being restricted to a few shreds in iron-oxide patches. Black iron-oxide is abundant in irregular grains, the borders of which occasionally form a vermicular intergrowth with plagioclase in orthoclase. Zircon is a common accessory. A similar rock, now much brecciated, comes from below the Red Rock capping Dunnideer. The rock seen in the quarry on the top of Fallow Hill may belong to this stage, but it is richer in biotite and hornblende than those just described. It contains a small amount of quartz wedged between the feldspars.

The final term in the transitional series is the Red Rock (Plate 18 A). As already mentioned, this rock is miarolitic, often brecciated and silicified, pale red in colour and shows large plates of pinkish feldspars and scanty greenish mafic patches. A typical representative from the top of Hill of Johnston is seen to be composed, in order of abundance, of orthoclase, oligoclase, amphibole, biotite, chlorite, quartz, apatite and zircon. The orthoclase, tabular perthitic and slightly sericitised, forms nearly half of the rock. Oligoclase, An_{20}, occurs in two forms—one small, lath-shaped, cloudy and saussuritised enclosed in the orthoclase and very much corroded, and the other broad, platy, fairly clear and with little replacement; small myrmekitic intergrowths are common. Quartz is present in intergrowths or it occupies a few small interspaces between feldspars. Mafic patches amount to about a third of the rock. They consist mainly of chloritised and modified spongy hornblende, which may itself be after a pyroxene. Biotite occurs in the mafic clots in ragged blades and aggregates. Prisms of apatite and zircon are markedly abundant in all the other constituents. The mode is orthoclase 48 %, oligoclase 17 %, quartz 0.7 %, mafic patches (amphibole, chlorite, biotite, iron-oxide) 34 %, apatite and zircon 1.2 %. Some Red Rocks show an increase in potash feldspar till plagioclase is almost absent. In others quartz may become fairly conspicuous or the rocks may show numerous cavities lined with iron-oxide.

Specific gravities of most of the analysed rocks are given in the Tables of chemical compositions and are consistent with their mineralogical compositions and place in the differentiated series.

(*iii*) *Chemical Composition of the Syeno-Gabbros and Red Rocks*. In Table II are given nineteen analyses of the syeno-gabbros and syenites from the Red Rock Hills. Fifteen of these analyses were done by Haq and four by Herdsman. They are projected on the triangular diagrams of Figs. 3 and 4. Though some of the analyses are of rather indifferent material, due to the often brecciated and carious nature of these late derivatives, nevertheless they confirm clearly the course of the primary differentiation indicated by the petrographic study. At this present, attention need be directed only to the following points. These rocks occupy a field in the $FeO+Fe_2O_3 : MgO : Na_2O+K_2O$ diagram towards the alkali-iron border, and have a consistent low value for MgO. From syeno-gabbro to Red Rock the rocks show a trend towards the alkali apex. In the lime-alkali diagram an enrichment in potash is indicated concordant with the observed enrichment in orthoclase.

The summations of the analyses in this Table, as in all the tables, leave margins for the trace elements whose distribution is to be dealt with in a later paper. The low summations in the syenites and some syeno-gabbros are in part due to the remarkable increase of the barium content in these late terms of the differentiated series. Haq has determined spectroscopically values for barium in these rocks ranging from about ½ % to 1½ %—by contrast an olivine-gabbro gave about .05 %. Presumably the barium is proxying for potassium in the potash-feldspars and micas in the late differentiates. This barium concentration is a matter for further investigation.

(c) Petrogenesis of the Insch Differentiated Suite

From the field, petrographic and chemical evidence presented in the foregoing, there can be no doubt that the olivine-gabbro, syeno-gabbro and syenite form together a differentiated suite. The question of the nature of the undifferentiated magma of the suite is one of some interest. The amount of the acid differentiate represented by the Red Rock syenite cannot be very large; the fact that the Red Rock does not appear along the

TABLE II: Syeno-Gabbros and Syenites

	20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38
SiO_2	44.7	42.9	45.6	50.3	51.47	49.6	51.36	52.0	55.68	53.2	56.9	56.8	59.18	53.2	59.3	50.9	51.0	58.4	59.0
TiO_2	2.5	2.6	2.3	1.8	1.33	1.9	1.07	1.8	0.88	1.7	1.0	1.1	0.68	1.4	0.9	1.3	1.10	1.10	0.77
Al_2O_3	14.9	17.1	17.2	13.0	16.86	14.2	19.62	15.3	20.14	15.9	17.8	16.6	18.26	16.5	17.1	18.5	22.6	16.6	19.3
Fe_2O_3	1.1	0.7	1.7	1.0	2.11	1.0	1.43	1.5	3.34	1.8	3.5	2.2	2.53	4.2	2.7	3.5	2.5	2.6	2.4
FeO	14.9	15.2	13.0	17.1	13.58	15.8	9.96	9.6	6.17	7.9	5.8	6.6	5.76	5.0	5.5	4.3	3.0	4.9	4.0
MnO	0.37	0.37	0.31	0.61	0.46	0.37	0.39	0.34	0.18	0.24	0.16	0.07	0.25	0.23	0.15	0.17	0.21	0.21	0.11
MgO	5.6	4.2	3.8	2.5	1.07	2.4	1.46	3.2	0.69	2.0	0.99	1.6	0.68	2.6	0.73	2.5	2.7	1.9	1.5
CaO	9.4	9.0	7.8	5.7	5.68	5.4	6.22	6.3	1.83	5.3	1.6	4.1	0.29	4.6	0.48	5.9	9.2	1.6	0.81
Na_2O	3.6	4.1	4.1	2.5	2.86	3.0	2.55	3.2	3.72	3.2	2.9	3.0	2.81	4.6	2.5	4.7	4.1	2.8	2.2
K_2O	1.0	1.8	2.0	3.5	3.28	3.4	4.37	3.2	4.03	4.6	5.8	5.0	6.26	4.1	7.4	4.0	1.4	6.5	6.5
P_2O_5	1.1	0.91	0.93	0.40	0.39	0.56	0.49	0.72	0.29	0.62	0.27	0.26	0.14	0.48	0.12	0.61	0.14	0.34	0.21
H_2O	0.60	0.53	0.59	0.80	0.72	0.92	0.82	1.7	2.80	1.5	1.8	1.6	3.20	1.9	2.2	1.4	1.4	2.3	2.1
Total	99.77	99.41	99.33	99.21	99.81	98.55	99.74	98.86	99.75	97.96	98.52	98.93	100.04	98.81	99.08	97.78	99.35	99.25	98.9
S.G.	3.00	3.15	3.02	3.10	—	3.01	—	2.91	—	2.84	—	2.60	—	2.68	—	2.80	2.74	2.72	2.62

northern contact suggests that it forms a sort of bubble or lens, the original extent of which was not much greater than an area enclosing the Red Rock Hills and its original thickness not much more than that of the Red Rock zone now exposed, namely some 200 feet; the intermediate syeno-gabbro zone is still thinner. When the great area occupied by the olivine-gabbro is recalled, it appears certain that the production of the more acid differentiates has not resulted in any marked increase in the basicity of the olivine-gabbro magma itself.

When compared with the three other analysed olivine-gabbros of the north-east Scottish Basic Province (Belhelvie, B in Fig. 3, Huntly, H, and Haddo House, HH), the Insch olivine-gabbros are seen to be much richer in iron-oxides. We may recall that the analysed Insch olivine-gabbros appear on the evidence at present available to become richer in iron-oxides as they approach the Red Rock Hills until they attain an iron-oxide content of some 24% and a specific gravity of 3.2. Whilst a general similarity exists between the course of the Insch differentiation and those of the Skaergaard liquid (Wager & Deer, 1939), the Duluth rocks (Grout, 1918) and Duluth magma (Cornwall, 1951), it is clear that the Insch olivine-gabbros are all nearer the iron-enriched stage. It may accordingly be concluded that the Insch olivine-gabbros represent a more advanced stage in the differentiation or iron-enrichment of the gabbro magma of the province. There should be somewhere a magnesium-enriched cumulate commensurate in size with the iron-enriched olivine-gabbro—the question of its whereabouts will be our concern in a later page. Whilst it is possible that in its progress towards iron-enrichment the Insch magma at first followed the Skaergaard pattern, later it became potash-enriched with the production of syenitic rocks, whereas the Skaergaard differentiation gave ferro-gabbroic rocks and a small fraction of granophyric type. When comparison is made with the trend in the Duluth gabbro a closer parallelism is apparent. Both Insch and Duluth ended up in the same kind of syenitic Red Rock, with the Insch examples being rather the richer in alkalies and iron. The series from olivine-gabbro to orthoclase-gabbro and finally Red Rock in the two complexes are remarkably alike. Both gabbro masses belong to the Duluth Type defined by Cooper (1936).

Some such process as the following may be proposed to have operated in the production of the Insch Differentiated Series. As the refractory components of the olivine-gabbro magma crystallised, the residual magma became enriched in the less refractory constituents, especially alkalis. The first stage of this enrichment is given by the orthoclase-rich gabbros and the final stage is represented by the materials for the syenitic Red Rock accumulated as the result of the gradual upward expulsion of the low-temperature residue from the settling crystal mush. The volume of this final term was small; it is maybe more comparable with that (0.15%

deduced by Cornwall (1951, 170) for the Greenstone Flow in the Keweenawan, than that (2%) suggested by Grout (1918, 639) for the Duluth Red Rock.

3. THE BOURTIE DUNITE–TROCTOLITE SERIES

We may now take up the description and discussion of the second unit of the Insch gabbro body, the basic and ultrabasic rocks developed at the east end of the Insch Mass around Bourtie.

(*i*) *Field Occurrence.* The field relations and petrography of these Bourtie rocks have been dealt with by Sadashivaiah (1954b) in his discussion of the form of the eastern part of the Insch Mass. The distribution of the outcrops (Sadashivaiah, 1954b, fig. 2) agrees with a large-scale banding of the rock-types trending about north and south, as indicated in our Fig. 1. The peridotite, troctolite and olivine-gabbro belts shown in this figure are partly diagrammatic and it might be preferable to view this portion of the Insch Mass simply as a banded ultrabasic and basic complex. Beginning at the eastern limit, we find the crude succession of the members of this complex to start with a broad belt of dunite and peridotite with subordinate bands of troctolite, as seen near Woodside and around the old fort on Barra Hill. Westwards comes a belt of olivine-rich troctolite and peridotite, exposed in the coppice due west of the fort, that shows a fairly definite north–south banding, apparently steeply inclined. Westwards again to Barra Castle, dunite-peridotite and peridotite are found in a couple of quarries. Immediately south of Barra Castle, large troctolite and peridotite blocks, possibly not much off place (cf. Blundell & Read, 1958), are abundant. West of Barra Castle great blocks of olivine-gabbro appearing in the flat ground are not definitely in place and are not farther considered here. Finally, a quarry at the road junction east of Mill of Bourtie exposes solid dunite. The Bourtie rocks certainly in place therefore present a banded series of dunite, peridotite and troctolitic gabbro.

(*ii*) *Petrography.* The ultrabasic rocks include both dunites and peridotites. These are greenish-black rocks speckled with a few small whitish areas. In slice, the dunites (Plate 18 B) are seen to be composed almost entirely of partly serpentinised straight-edged olivine grains, usually closely packed together but in places separated by angular areas of dense 'bastitic' material presumably derived from a pyroxene. In the peridotites, plates of orthopyroxene of Bushveld type (Hess & Phillips, 1938) occur, more or less altered but showing fine lamellae of clinopyroxene; they enclose small rounded grains of olivine. Occasionally, large independent euhedra of orthopyroxene of Bushveld type appear between the olivines. Small tabular plates of clinopyroxene are also found to have been partly transformed into bastite. Relics of poorly developed double coronas around olivine have been noted and indicate that plagioclase originally filled some of the interstices between the olivine crystals—but such feldspars have completely altered to an isotropic amorphous grey mass. Haq determined the olivine as having a forsteritic composition of Fo_{78}.

By the increase of pyroxene and feldspars in definite streaks and bands, the ultrabasic rocks pass into troctolitic peridotites and gabbros. The white specks increase in size and number into irregular bands. The rocks in hand specimens look like troctolites, although slices show that they have the simple troctolite composition only in narrow bands. The main components in order of abundance are plagioclase, pyroxenes and olivine, with very minor deep brown hornblende and red-brown biotite. An average plagioclase is a labradorite, An_{63}, in tabular crystals showing various degrees of alteration. By far the most important pyroxene is an orthorhombic variety (En_{72} in one example) forming large euhedra, often clustered and usually showing delicate striping with clinopyroxene. Large ophitic plates of clinopyroxene are patchily distributed in some rocks. Olivine forms large irregular serpentinised grains, blackened with iron-oxide and in some rocks orientated, either by flow or by sedimentation. Beautiful double coronas of the type described by Read (1923, 106, 107, fig. 7) from gabbros of the Huntly Mass and by Stewart (1946, 472) from Belhelvie, are developed around the olivines at their contacts with feldspar: the narrow inner rim is usually of orthopyroxene and the broader outer rim of fibrous amphibole, often symplectic. Partially destroyed coronas enclosing olivine undergoing various degrees of serpentinisation show that this process occurs later than the formation of the coronas. One of the bands of basic rock in the coppice west of Barra Hill fort lacks the troctolitic appearance of the rocks described above and the thin slice shows that it is poorer in olivine and richer in orthopyroxene; the latter is as abundant as the plagioclase and is often enclosed as separate euhedra in large plates of clinopyroxene (Sadashivaiah, 1954b).

(*iii*) *Chemical Composition.* In Table III there are given twelve analyses of Bourtie rocks made by Haq. The analyses of the ultrabasic and near-ultrabasic rocks have been recalculated to water-free weight-percentages to a total as near as reasonable to that of the original analyses. The chemical compositions of the rocks are what is to be expected from their mineral composition and necessitate few comments at this stage. In the ultrabasic rocks, the low SiO_2, TiO_2, Al_2O_3, CaO, Na_2O and K_2O and high MgO are compatible with a forsteritic olivine as their main mineral component. The ratio of total iron to magnesium is found to be practically the same for all the rocks, showing that the factor for precipitation of Fe and Mg is nearly constant.

The small-scale rhythmic banding in the Bourtie rocks is similar to that described from many great gabbroic intrusions such as Bushveld, Duluth, Stillwater, Bay of Islands, Skaergaard and others. The nature of the Insch

TABLE III. Bourtie Dunite–Troctolite Series

	DUNITES								DUNITE–PERIDOTITE	
	1a	1b	2a	2b	3a	3b	4a	4b	5a	5b
SiO_2	36.2	41.6	38.0	42.8	35.8	40.8	38.1	42.6	35.4	40.2
TiO_2	0.05	0.06	0.06	0.07	0.06	0.07	0.12	0.13	0.10	0.11
Al_2O_3	3.0	3.3	3.2	3.7	3.3	3.8	6.2	6.9	4.8	5.4
Fe_2O_3	10.4	12.0	10.3	12.4	8.9	10.15	6.9	7.7	5.5	6.2
FeO	2.9	3.3	1.8	2.1	4.0	4.6	2.5	2.8	5.2	5.9
MnO	0.19	0.22	0.19	0.22	0.19	0.22	0.15	0.17	0.17	0.19
MgO	34.0	39.0	32.1	37.0	34.3	39.14	33.6	37.6	35.5	40.3
CaO	0.14	0.16	0.29	0.33	0.14	0.16	1.60	1.8	0.75	0.85
Na_2O	0.08	0.09	0.14	0.16	0.10	0.11	0.12	0.13	0.10	0.11
K_2O	0.06	0.07	0.06	0.07	0.07	0.08	0.09	0.10	0.07	0.08
P_2O_5	tr.	—	0.04	0.05	0.04	0.05	0.02	0.02	tr.	—
H_2O	12.4	—	12.7	—	12.3	—	10.4	—	11.8	—
Total	99.42	99.80	99.38	99.90	99.20	99.18	99.80	99.95	99.39	99.34
S.G.	2.69		2.67		2.63		2.64		2.42	
$\frac{FeO+Fe_2O_3\times 100}{FeO+Fe_2O_3+MgO}$		28.9		28.1		27.4		21.8		23.1

	PERIDOTITES								OLIVINE-RICH TROCTOLITE		HYPERSTHENE-RICH OLIVINE-RICH GABBRO		TROCTOLITE
	6a	6b	7a	7b	8a	8b	9a	9b	10a	10b	11a	11b	12
SiO_2	37.5	41.7	36.1	40.9	37.4	42.1	35.0	38.6	38.0	41.9	41.8	45.3	47.2
TiO_2	0.10	0.11	0.07	0.08	0.11	0.12	0.11	0.12	0.09	0.10	0.28	0.30	0.60
Al_2O_3	3.0	3.3	7.3	8.3	7.7	8.7	1.7	1.9	5.0	5.5	9.4	10.2	12.7
Fe_2O_3	7.8	8.7	5.8	6.6	7.3	8.1	12.2	13.5	4.0	4.4	3.8	4.1	1.4
FeO	4.6	5.1	4.7	5.3	4.0	4.5	4.0	4.4	3.6	4.0	5.2	5.6	7.5
MnO	0.18	0.20	0.14	0.16	0.15	0.17	0.16	0.18	0.12	0.13	0.16	0.17	0.14
MgO	33.6	37.4	29.4	33.3	28.0	31.6	34.3	38.0	33.2	36.6	25.8	27.9	18.1
CaO	2.3	2.6	3.8	4.3	3.0	3.4	2.4	2.7	5.5	6.1	4.4	4.8	7.9
Na_2O	0.19	0.21	0.15	0.17	0.16	0.18	0.17	0.19	0.89	0.99	0.77	0.83	1.8
K_2O	0.07	0.08	0.07	0.08	0.08	0.09	0.07	0.08	0.09	0.10	0.15	0.16	0.37
P_2O_5	0.05	0.06	0.04	0.05	0.04	0.05	0.04	0.04	0.03	0.04	tr.	—	0.06
H_2O	10.0	—	11.7	—	10.9	—	9.3	—	9.3	—	7.6	—	1.7
Total	99.39	99.46	99.27	99.24	98.74	99.01	99.45	99.71	99.82	99.86	99.36	99.36	99.47
S.G.	2.75		2.71		2.73		2.74		2.74		2.89		2.96
$\frac{FeO+Fe_2O_3\times 100}{FeO+Fe_2O_3+MgO}$		26.9		26.1		28.5		32.0		18.7		25.8	32.9

exposures does not warrant any contribution to the discussion of the actual mechanism of the production of this banding. We can, however, state our preferences; from what is seen, it appears unlikely that the diversity of this unit arises either from successive intrusions or from the intrusion of an already differentiated pasty magma. After comparison with well-exposed investigated gabbroic bodies, we prefer to regard the Bourtie unit as an example of gravitative differentiation in place. Accumulations of the more refractory minerals, rich in magnesia, separating out early from an initial magma led to the formation of the *cumulates*, dunite, peridotite and troctolite. The ultrabasic rocks include *orthocumulates* in the nomenclature of Wager, Brown & Wadsworth (1960).

If this origin of the Bourtie basic and ultrabasic rocks by the accumulation of early formed magnesia-rich minerals is admitted, then the question immediately arises as to whether the parent magma from which these minerals came is represented in the Insch area. Put another way, are the Bourtie magnesia-rich cumulates derived from the same gabbro magma that gave rise to the iron-enriched olivine-gabbros of the Insch Differentiated Suite? Is there in the Insch Mass one grand differentiated series varying from dunite to syenite? We consider this question and possible ways of answering it in the next and last section of this account.

4. A POSSIBLE WIDER DIFFERENTIATED SUITE: DUNITE TO SYENITE

The Bourtie outcrops are not connected by visible exposures to the olivine-gabbros occurring to the west and making the main part of the Insch Differentiated Suite. It is true that there are bands in the Bourtie series that are in some ways like the olivine-gabbros and that olivine-feldspar-rocks are found within the main body of the olivine-gabbro, though, it is also true, such rocks are not identical with the Bourtie troctolites. It is true, too, that appropriate rocks and their minerals of the two units are petrographically very similar. It would not be unreasonable in this kind of ground to view the Bourtie Series as magnesia-rich cumulates complementary to the iron-rich olivine-gabbro. But, while all this may be so, it must be admitted that an unexposed gap of over a mile separates the exposed peridotites of Mill of Bourtie from the exposed olivine-gabbro of Cuttlecraigs.

In spite of these uncertainties, we are nevertheless inclined to the opinion that we are in fact dealing with a wider differentiated series, ranging from dunite to syenite. Because of this opinion, we have numbered the analyses consecutively from east to west, from the lowest cumulate at Bourtie to the highest syenite of the Red Rock Hills, and have plotted them all in the diagrams of Figs. 3 and 4.

In an earlier page, it has been suggested that the olivine-gabbro itself shows variations in its mineral phases from east to west up to the Red Rock Hills. Certainly, the iron-enrichment increases in the same direction. The detailed investigation of certain mineral phases of the olivine-gabbro and their comparison with corresponding phases of the Bourtie rocks might establish a reasonably complete series connecting the two units. It is proposed to begin a chemical and mineralogical investigation of the olivines and feldspars of strategically placed olivine-gabbros, troctolites and ultrabasic rocks to test the validity of the wider differentiated suite. (The separation of the two pyroxenes of these rocks appears to be a matter of great difficulty.) Further, results of some value might ensue from a study of the trace-elements of significant rocks; this aspect is receiving attention.

Meanwhile, we advance the provisional conclusion that crystal-fractionation in place of a gabbro magma of the Insch Mass resulted in, first, an enrichment of iron with respect to magnesium—this latter being removed in the Bourtie cumulates—and, next, an enrichment in alkalies, especially potassium, and silica. If this conclusion is correct then certain interesting speculations can be made. The volume of the iron-enriched olivine-gabbro is so great that the complementary magnesium-enriched cumulates must also be of considerable size. The interpretation, already referred to, of the natural contacts of the Insch Mass as roof means that the floor of the wider differentiated series is not visible. The detection and location of such an ultrabasic body should be capable of achievement by geophysical methods. A close and careful gravimetric survey of the environs of the eastern end of the Insch Mass might produce remarkable results. Further, by the extension of such surveys we could explore the possibility that the exposed Insch Gabbros, big though they are by Scottish standards, are only a small part of a very much bigger body—maybe a Scottish 'Bushveld'. We are approaching here the suggestion, made long ago by Read on quite unsound evidence, that the exposed basic masses of the north-east Scottish Province might make one huge sheet.

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APPENDIX: LOCALITIES OF ANALYSED ROCKS

1. Dunite, west slope of Hill of Barra, Bourtie.
2. Dunite, west slope of Hill of Barra, Bourtie.
3. Dunite, west slope of Hill of Barra, Bourtie.
4. Dunite, old quarry, at road junction, 500 yards east of Mill of Bourtie.
5. Dunite-peridotite, old quarry, 220 yards east of Barra Castle, Bourtie.
6. Peridotite, west slope of Hill of Barra, below Fort.
7. Peridotite, chevron coppice, 800 yards east of Barra Castle, Bourtie.
8. Peridotite, 500 yards south-east of Barra Castle.
9. Peridotite, 500 yards east of Barra Castle.
10. Olivine-rich troctolite, chevron coppice, 800 yards east of Barra Castle.
11. Hypersthene-rich olivine-rich gabbro (block), road-fork, quarter-mile south of Barra Castle.
12. Troctolite, south slope of Hill of Barra, immediately below Fort.
13. Olivine-gabbro, one-third mile east of church, Kirktown of Rayne.
14. Olivine-gabbro, old quarry, Bonnyton, 3 miles north-east of Insch.
15. Olivine-gabbro, same locality as No. 14. Analyst, E. G. Radley, *in* Read, 1923, p. 124.

16. Olivine-gabbro, old sand-pit, South Lodge of Freefield House, two and three-quarter miles north-east of Insch.
17. Olivine-gabbro, sand-pit, Old Gateside, two and one-third miles north-north-east of Insch.
18. Olivine-gabbro, 220 yards south of summit of Hill of Newleslie, three and a half miles south-west of Insch.
19. Olivine-gabbro, 100 yards south of the summit of Hill of Johnston, four miles south-west of Insch.
20. Syeno-gabbro, 76 yards east-south-east of summit of Hill of Newleslie.
21. Syeno-gabbro, 90 yards south-south-east of summit of Hill of Johnston.
22. Syeno-gabbro, 95 yards south-east of summit of Hill of Johnston.
23. Syeno-gabbro, high up on east face of quarry at south-west end of Hill of Johnston.
24. Syeno-gabbro, quarry as No. 23. Analyst, Herdsman.
25. Syeno-gabbro, lowest part of south face, quarry as No. 23.
26. Syeno-gabbro, roadside quarry, south-west base of Gallow Hill, two and one-third miles south-west of Insch. Analyst, Herdsman.
27. Syeno-gabbro, 70 yards south-south-east of summit of Hill of Johnston.
28. Syenitic Red Rock, summit of Hill of Johnston. Analyst, Herdsman.
29. Syenitic Red Rock, near No. 25, quarry, Hill of Johnston.
30. Syenitic Red Rock, summit of Hill of Johnston.
31. Syeno-gabbro, quarry, south-west base of Gallow Hill.
32. Syenitic Red Rock, summit of Gallow Hill. Analyst, Herdsman.
33. Syenitic Red Rock, Summit of Hill of Dunnideer, one and a quarter mile west of Insch.
34. Syenitic Red Rock, near summit of Gallow Hill.
35. Syenitic Red Rock, 150 yards south of the summit of Hill of Dunnideer.
36. Syeno-gabbro, 120 yards south of the summit of Hill of Newleslie.
37. Syenitic Red Rock, 300 yards south-west of the summit of Gallow Hill.
38. Syenitic Red Rock, 420 yards south-west of the summit of Gallow Hill.

(Unless otherwise stated, analyses are by B. T. Haq.)

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River Terraces and Drainage Development in the Reading Area

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ABSTRACT: An examination is made of the terrace succession in the middle and lower reaches of the Rivers Kennet and Loddon–Blackwater and of the area around their junctions with the Thames between Pangbourne and Wargrave. By a critical examination of the altitudinal and field relationships of these terraces it is found possible to indicate correlations with the terrace succession of the main stream established by F. K. Hare and K. R. and Catherine E. Sealy. In the light of the new evidence presented, a re-assessment is made of the current theories for the mechanisms of drainage diversions in the Reading area.

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1. INTRODUCTION

Reading lies at a nodal position in the London Basin, below the Goring Gap through which the Thames enters from the north to join the longitudinal consequent system comprised of the Kennet and the middle and lower reaches of the Thames itself. Below this confluence the Loddon–Blackwater system, draining the country to the south and east of Reading, finds its way to the Thames at Wargrave.

Details of the terrace record left by the middle Thames have been traced from Slough to the neighbourhood of Reading by Professor F. K. Hare (1947) and by Dr. and Mrs. K. R. Sealy (1956). The logical extension of these studies lies in the elucidation of the terrace succession in the valley of the River Kennet, since this stream rather than the upper Thames is the lineal relation of the trunk stream below Reading. The Goring Gap seems to mark an apparent discontinuity in the terrace record between the upper and middle reaches of the Thames. The major changes in the drainage patterns of this region of convergent tributaries have affected not only the courses of the middle Thames and the lower Kennet, but also those of the Loddon and the Blackwater.

The following account is therefore devoted to a consideration of the surviving remnants of the terrace sequences in the middle and lower reaches of the Rivers Kennet, Loddon and Blackwater and also of some of their tributary streams. Over this rather dissected area it has been possible to map only the terrace surfaces; the resultant map (Fig. 1) is thus directly comparable with that produced for the middle Thames by Dr. and Mrs. Sealy.

The field survey was undertaken on a scale of 1:25,000, using the larger 1:10,560 scale maps for reference. The determination of heights was achieved with an aneroid altimeter of the 'Paulin' type with which it was found possible to register heights within a range of accuracy of ten feet, corroborating the now well-substantiated claims of other workers using this method of survey. The method of using the altimeter to obtain this order of accuracy has varied from one surveyor to another. Whereas B. W. Sparks by restricting his traverses to one hour in duration, and taking care to select suitable occasions for the surveying, considered it sufficient to distribute the error of closure at the end of each traverse, Dr. and Mrs. Sealy adopted a method, elaborated in their joint paper, which attempts to eliminate the factor of time from the readings, since it is the variation of meteorological elements with time which is responsible for the more serious errors in aneroid readings. This method is, however, somewhat slow and requires rigorous attention to timing. In the survey undertaken here it was found possible to apply corrections to field readings by reference to meteorological records kept at the University of Reading. Thus while using basically Sparks' method, the time of each reading was noted and a correction subsequently applied to each figure for pressure change and for temperature. Since variations in these elements may be compensating and not cumulative in their effect on the readings, the simple distribution of errors of closure may in certain circumstances add to the residual error in the readings.

Height readings were taken on the rear and leading edges of each flat and at median points where necessary. The terrace remnants have been mapped as morphological features and include what Sealy refers to as, 'slightly degraded terrace forms'. This has been done not only to include significant features which would otherwise have been excluded, but also because of the difficulty of recognising and delimiting 'terminal surfaces'.

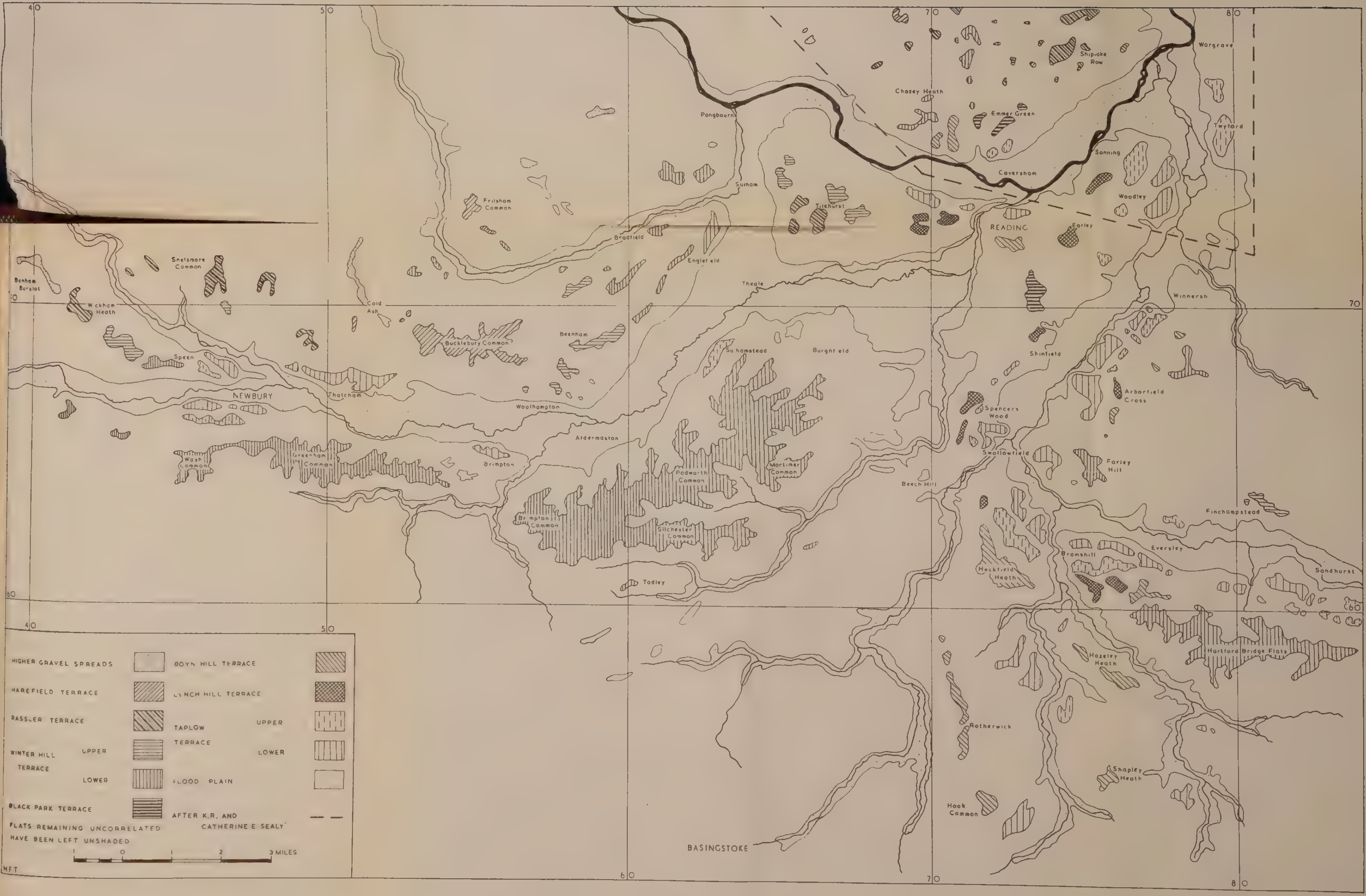

Fig. 1. The terraces of the Kennet and Blackwater drainage systems

The eye can map terrace margins only by reference to significant breaks of slope which are genetically related to post-depositional weathering of the terrace and not to the terminal surface, the delimitation of which is a highly subjective matter. Apart from this inherent difficulty, from the geographical point of view, flats are more appropriately mapped according to purely morphological criteria, providing always that the height determinations are made within the boundary of the terminal surface, where this is sensibly preserved.

The original method followed in the analysis of the information presented in this study was to consider the Kennet and Loddon–Blackwater systems as separate entities, and the terraces recognised were given only local significance. But by tracing each terrace downstream to the present or past confluence with the River Thames it has been found possible to relate the tributary terrace sequence in each case to the established Thames succession following the work of Professors S. W. Wooldridge and F. K. Hare and Dr. K. R. and Mrs. C. E. Sealy.

2. THE TERRACES OF THE KENNET SYSTEM

Although essentially a longitudinal consequent stream, the River Kennet over most of its course is not fully adjusted to structure, nor are the northern tributaries, the Lambourn and the Pang, which formerly flowed into the Kennet but now reaches the Thames at Pangbourne, simple dip-slope consequent streams (Fig. 2). The Kennet flows to the north of the main axis of the London Basin and, east of Aldermaston, obliquely against the dip. The course of the Lambourn appears to be guided by slight, local flexures of post-Eocene age, mapped by Beynon in an unpublished study of the Berkshire Downs. The Pang begins its course as a consequent

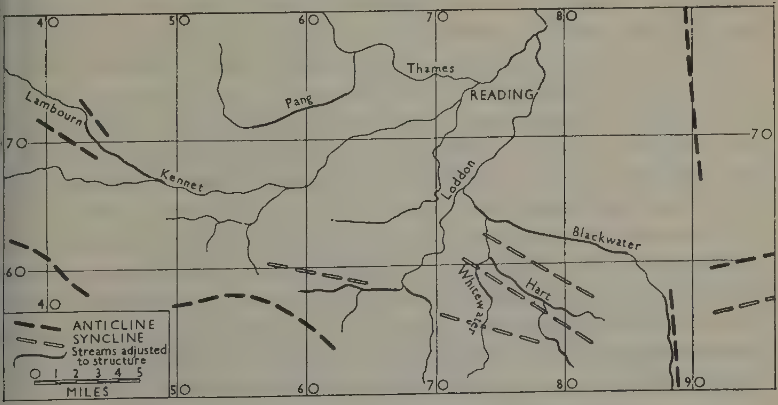

Fig. 2. Streams and structures in the Reading area

stream; it was probably originally part of the late Tertiary consequent system and joined to the Windrush, now a tributary of the upper Thames. The middle reaches of this small but picturesque river have become diverted, possibly by capture, into a course parallel to the strike of the chalk, while the lower portion of the stream survives as a remnant of the former obsequent course of the Kennet.

The recognition of the principal high terraces in the Kennet Valley was achieved as early as 1907 by Osborne White, who distinguished three stages in the denudation of the Kennet Valley:

(i) The Cold Ash Stage, 265 feet above the flood-plain,
(ii) The Bucklebury Stage, 230 feet above the flood-plain,
(iii) The Silchester Stage, 155 feet above the flood-plain.

Both White and later Hawkins have made references to the lower terraces well represented in the neighbourhood of Newbury.

The following account refers where appropriate to the stages recognised by White, but in order to simplify the presentation of the material each of the terraces which have been traced in this study is discussed under the heading of the appropriate Thames terrace with which it has been correlated.

(a) High-Level Features

The highest recognisable terrace of the proto-Kennet is represented by the long narrow gravel capped outlier of the Lower Bagshot Beds stretching northwards from Cold Ash (513696),[1] and by a comparable flat farther west near Benham Burslot (406707). Six miles east of Cold Ash a small remnant near Upper Basildon (597763), standing at 468 feet O.D. and underlain by gravels containing diagnostic northern drift materials, may well represent the same early stage in the present cycle near the Kennet–Thames confluence. These are probably scattered remnants of an early post-Pliocene course of the Kennet which then flowed near to the axis of the Pliocene trough but high up on the northern limb of the mid-Tertiary syncline. This terrace, if such it can be called, is the Cold Ash Stage of H. J. O. White, but correlation with the higher flats bordering the Caversham–Henley channel of the Thames is only possible to the extent that these remnants can be assigned to the Higher Gravel Train of S. W. Wooldridge. Adjustment to the structural trends in the region began early in the current cycle of erosion, and is manifest primarily in the process of uniclinal shifting of the Kennet–Thames. Since the Pliocene feature and the earlier structure diverge eastwards this process has been more marked in the lower reaches of the system. It is, however, evident in the early stages of development of the River Kennet. But perhaps the dominant feature of stream migration in the whole area under consideration has been the

[1] All grid references lie within the 100 km. square 41 (SU).

tendency for tributary streams to defer confluence with the parent river. This feature, which Osborne White described as 'postponed junction' (1902), is an expression of the greater velocity and load of the master stream. This process, continued during a series of stages in the denudation of the present valley, has produced composite interfluves between subparallel streams preserving terrace remnants in 'longitudinal sequence', distinct from the stairway of terraces in 'parallel sequence', resulting from uniclinal shifting. Thus in the Kennet basin successive terraces are found preserved not only southwards but also eastwards from the highest features.

(b) The Harefield Terrace

The chief representative of this next major stage is found to the south-east of Cold Ash, as the gravel-covered plateau of Bucklebury Common which falls in height eastwards from 441 feet O.D. to 425 feet O.D. in just under two miles. The gradient of ten feet per mile maintains the terrace at a fairly constant height of 230 feet above the present flood-plain (Fig. 3). The gravels are usually from five to seven feet in depth and well stratified with frequent seams of clay. The land rises westwards to the Cold Ash level and flats can be recognised as continuing this terrace into the Pang Valley and upstream in the Kennet Valley near Ashmore Green (505696). Downstream of Bucklebury Common two flats, at 426 feet O.D. and 411 feet O.D., around Frilsham Common (546725), probably fall within this terrace, possibly representing the Pang–Kennet confluence at this time. This, the Bucklebury Stage of H. J. O. White, clearly represents a major phase of planation and deposition in the Kennet Valley. Its relationship with the Thames flats north of Reading is therefore of importance. Unfortunately

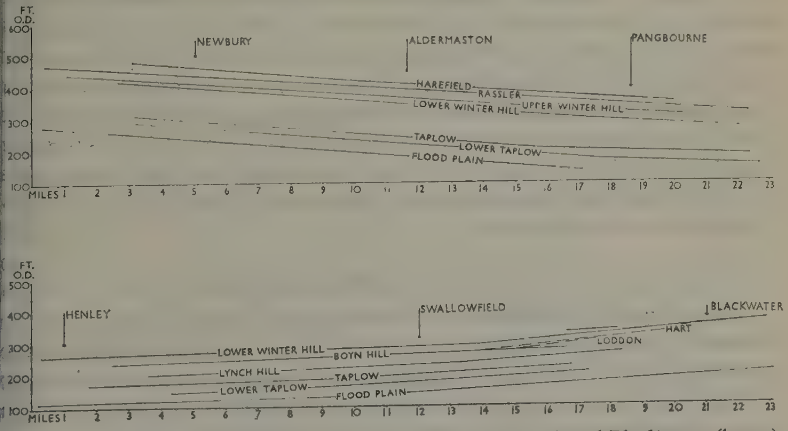

Fig. 3. The longitudinal profiles of the Rivers Kennet (upper) and Blackwater (lower)

east of Bucklebury Common itself the terrace is not represented in the Kennet Valley and correlation with the Thames sequence becomes a matter of speculation. A gradient of eight feet per mile, however, would bring this terrace into the Harefield of the Thames which is more probable than a slope of fourteen feet per mile which would relate it to the Rassler terrace (below), and by its field relationship with the Cold Ash gravels it is clearly a later stage than that represented by the High Gravel Train.

(c) The Rassler Terrace

Three flats standing some 200 feet above the present river, between 460 feet O.D. and 465 feet O.D., at Wickham Heath (417697), and two within the Lambourn Valley itself, Hoar Hill (434712) and Boxford Common (440715), appear to represent a lower stage than the Harefield terrace with which they could be correlated only if the upstream gradient from Bucklebury Common were assumed to be reduced from ten to five feet per mile. This is deemed improbable. Downstream the terrace is probably continued by Snelsmore East Common (480710) and Snelsmore Common (460710), which stand at heights between 458 feet and 444 feet O.D. These flats appear to represent a northerly swing of the Kennet over the present lower Lambourn Valley, from which the river has shifted uniclinally to its present position, while the Lambourn during this process has deferred confluence with the River Kennet. Together with the small flat south-west of Ashmore Green, at 439 feet O.D. these remnants suggest a downstream gradient of about eight feet per mile which, if maintained, would grade into the higher gravels above Tilehurst. These are found near Little Heath (655735) at heights between 340 feet and 345 feet O.D., some 200 feet above the Kennet but 215 feet above the Thames to which they are related since they contain materials of northern drift. They appear to represent a former junction with the Kennet, for, in the gravels of the most southerly of the three flats the northern constituents have been diluted with large amounts of flint. Dr. and Mrs. Sealy assigned these gravels to the Rassler terrace of the Thames, found downstream at Binfield Heath (743785). The only upstream representatives of such a terrace must be the gravels at Wickham Heath and Snelsmore Common. To relate the Little Heath flats to a higher or lower stage would require the somewhat improbable gradients of either fourteen feet per mile or nil.

(d) The Upper Winter Hill Terrace

Well below the Wickham gravels but higher than Osborne White's Silchester terrace, fragments of an intermediate level are well preserved near the village of Beenham (585688). A fairly extensive flat here falls from 351 to 338 feet O.D. in a little less than two miles. It lies about 170 feet above the flood-plain and more than fifteen feet above the adjacent

Silchester level. The field relationship and careful extrapolation do not bear out Mrs. Sealy's tentative suggestion that this forms the northern margin of the Silchester flood-plain. As a separate terrace it is continued upstream in small flats north of Woolhampton (574668) at around 360 feet O.D. and probably at Stockcross (435685) at 436 feet O.D. Downstream there is again a region of discontinuity towards Reading but remnants of a former southerly swing of the River Thames are found between 312 feet and 306 feet O.D. on the northern edge of the Tilehurst plateau, and a continued slope of seven feet per mile would grade the flats around Beenham into this terrace. Such an assumption is supported by the field relationships of the Beenham gravels with the Silchester gravels and by the fact that the latter, sloping at seven feet per mile, could not grade into the Tilehurst flats, even allowing for a substantial flattening towards the Thames confluence. Examining the evidence downstream Mrs. Sealy correlated these remnants with the Upper Winter Hill stage of the Thames. The evidence upstream is consistent with this view.

(e) The Lower Winter Hill Terrace

Osborne White (1907) emphasised the importance of this terrace when he wrote that the present relief of this whole region south of Reading has been 'carved out of a drift mantled plain of sub-aerial erosion whose gentle slopes converged like a shallow funnel somewhere in the vicinity of Reading'. This generalisation is probably correct, but the remnants of this plain do not differ in form from the higher or lower terraces although they are generally of greater extent. However, unlike the higher terraces of the Kennet, the main representatives of this stage are found to the south of the present river, marking the last stage when uniclinal shifting was an important influence in the shaping of the valley.

Two extensive plateaus stretch east and west either side of the Enborne Valley near Brimpton (558648). Upstream, Crookham, Greenham and Wash Commons fall from 412 feet O.D. to 372 feet O.D., while downstream the terrace is continued in the Brimpton, Silchester, Padworth and Burghfield Commons, falling from 356 feet O.D. to 312 feet O.D. The longitudinal gradient taken over the whole length of over twelve miles shows a slight flattening downstream from ten to seven feet per mile (Fig. 3). Other remnants bearing a consistent relationship with these larger plateaus occur along the northern flank of the valley. One west of Speen (455682) at 411 feet O.D., is one of a series of flats in longitudinal sequence now forming the interfluve between the Kennet and the Lambourn and presumably representing the confluence of the two streams at each respective stage. Downstream in a similar position along the Kennet–Pang interfluve are a number of flats scattered over the rather dissected country between Beenham and Bradfield (604726). Mrs. Sealy considered these to be too low to preserve any terminal surfaces, but although seldom more than 200 yards in width their form is not markedly degraded and up to twelve feet of well-stratified gravel is preserved in a pit on May Ridge (606702). The largest of these remnants falls from 312 feet O.D. near South End (595704) to 303 feet O.D. at Bradfield Institution (605716), correlating well with the height of the main gravel spreads across the valley at Sulhamstead Abbots (647678), where the plateau stands at 311 feet O.D. If this correlation is correct then the gravels farther downstream, at 296 feet O.D. near Bradfield (605726) and across the Pang Valley near Maiden Hatch (620742) at a height of 284 feet O.D., continue the Silchester terrace towards the Thames confluence and provide the nearest link with the Winter Hill terrace represented south of Chazey Heath (695774), at a height of 271 feet O.D. It seems probable that the River Pang joined the Kennet in the area marked by the gravels west of Tutt's Clump (587712) while the Kennet itself flowed over the line of the present middle Pang.

From the many descriptions of the gravels associated with this terrace to be found in the early geological literature their composition and appearance appear to be remarkably consistent. Blake (1903) describes the gravels as being composed of 'sub-angular flints, much weather worn and broken, and of flint pebbles together with a few blocks of Sarsen', material entirely of local provenance, from the Chalk or the Tertiaries. However, though the composition of the gravels varies little, their lithology is not as uniform as the earlier descriptions suggest. South-east of Padworth Common the more characteristic stony gravel is to be seen in sections of over 200 yards in length (613636). But even here some areas are noticeably more sandy and thin lenses of sand are not uncommon. At the base of the deposit a coarse sandy bed seldom more than one foot thick appears more or less continuous and indicates the earliest phase of aggradation. On Mays Ridge (605700) another extensive section displays a similar deposit, but it is interrupted in one place by a more important seam of sand at least five feet in depth and twenty yards across. Sand and clay seams are also common in a pit off Dark Lane (615745) and the gravels are well displayed in large pits north of Ufton Nervet (635680) and near Sulhamstead Abbots (647683). Where the deposit is diversified with sand in the upper part contortions and festooning occur as they do in the gravels of the Hartford Bridge Flats (see below, p. 427). Elsewhere the uniformity and angularity of the debris render any structures in the gravels less visible and they are often apparently absent. However, from the marked festooning seen in many sections of this and lower terraces, all these gravels, above the Taplow level, would appear to have been subject to frost heaving, as one would expect if periglacial conditions are to be associated with the deposition of the gravels of each terrace.

There are significant variations in the depth of the gravel. On Wash

Common in the west it attains twenty feet in depth but thins progressively downstream to fifteen feet on Greenham Common, ten feet on Brimpton Common, and farther east to as little as six to eight feet. But it deepens towards the eastern margins near Sulhamstead Abbots to ten feet and across the valley along the northern interfluve there is twelve feet of sometimes sandy gravel on May Ridge and over fourteen feet in the pit off Dark Lane (615745). Such variations, together with the character of the gravels and the great extent of this terrace and of the related flats of the Blackwater Valley (see below, p. 426), indicate a considerable period of planation at this stage followed by a short rapid period of deposition, during which the gravels were laid down as the channel deposits of a network of fast-flowing braided streams. Evidence is now coming to light of the former existence of such channels on the surface of the deposit, through the work of Mr. R. Jarvis of the Soil Survey.

All these features and the steep gradient of the terrace are indicative of aggradation under conditions of a high ratio of load to volume. The precise relationship of this process to the vicissitudes of the Pleistocene climates must remain in doubt, but the planation associated with these gravels must have been interglacial while the gravels themselves suggest the products of solifluxion from the surrounding hill slopes. It is possible that some of this material did not reach the stream courses until the periglacial climate had given way to warmer conditions, as suggested by Arkell (1947). But it is felt that the aggradation cannot have been interglacial in the full sense of that term and that the ice sheets were still supplying meltwater into the upper Thames, for this river in the Winter Hill channel between Caversham and Henley had a very low gradient of around one foot per mile and at the same time deposited gravels comparable in depth to those found in the Kennet and Blackwater Valleys. A very large volume of water must have been flowing through this channel and it is only from the melting of northern ice that this could have been supplied, during an amelioration of climate which heralded its retreat.

When downcutting began after this prolonged period of planation and deposition, the river entrenched itself along the northern margin of its flood-plain, reversing the tendency for uniclinal shift. The reason for this is not clear, but the tendency was never very marked in this reach of the river and possibly the most important influence was the existence of more permanent channels, in the otherwise ever-changing braided patterns, along the northern edge of the plain coming into the middle Kennet Valley through the Lambourn, which from the network of dry valleys can be seen to have a large catchment area. Perhaps the course of the middle Pang also resulted from the confirmation of formerly braided channels where the river had deferred confluence during the Silchester aggradation.

Since this post-Silchester entrenchment, the River Kennet has worked and reworked over the same areas without any significant diminution in volume until recent times. Consequently few remnants of the later terraces have survived the successive periods of erosion and planation at still lower levels. Two small flats occur some 135 feet above the flood-plain, one at Midgham Church (555673) and the other on the Kennet–Enborne divide west of Brimpton (557647). Similar remnants occur north-east of Woolhampton (574668), between 90 feet and 110 feet above the flood-plain. Perhaps here are to be found the only surviving traces of the Boyn Hill and Lynch Hill terraces in the Kennet Valley but none is sufficiently extensive to justify correlation with these terraces.

At much lower altitudes, however, considerable areas are covered by later terrace gravels between sixty and twenty-five feet above the flood-plain.

(f) The Taplow Terraces

Between Newbury and Speen (455680) the spur between the Lambourn and the Kennet is bevelled at 306 feet O.D., fifty-six feet above the flood-plain of these streams. Some ten miles downstream a gravel-covered valley-side bench around Ufton Green (626683) stands fifty feet above the Kennet. These flats appear to represent a terrace fifty to sixty feet above the flood-plain, a height familiar in the Thames Valley, where it is recognised as the Taplow terrace. This terrace Dr. and Mrs. Sealy found to be composite in the Reading area. The same complexity is found in the Kennet and the Blackwater Valleys.

The most extensive representative of the lower flats extends westwards for about a mile and a half from Thatcham village (515674), itself largely built on this feature. Its height above the present stream is about thirty feet, and further remnants of this stage are found around Brimpton and Newbury itself, where south of the river the feature is, locally, composite, slightly lower flats apparently having been formed some twenty to twenty-five feet above the flood-plain. The lower terrace recurs again in the Sulham Gap, north of Englefield (625722) at 171 feet O.D., about twelve feet above the floor of the abandoned channel. Osborne White (1907) has described these gravels, from a section near Newbury station, no longer visible, as consisting 'almost exclusively of flints and sarsens embedded in seams of greenish clayey sand'. Such indications of torrential conditions of deposition, together with the height relations of the gravels, suggested to both Hawkins and Treacher correlation with the Taplow terrace of the Thames.

(g) The Flood-Plain

Professor Hawkins has shown that this is a composite feature containing a low terrace some six to ten feet above the present alluvium. The character of the present flood-plain deposits cannot here be adequately described or explained, but peat and marl above the gravels suggest recent shrinkage of the river both in volume and in load, and that gravel shoals of the earlier torrential conditions later became barriers to the enfeebled stream, and that lacustrine, and later swamp, conditions existed over wide areas of the valley. Arkell (1947), however, attributes some of these features to beaver dams of Bronze Age date.

3. THE TERRACES OF THE LODDON–BLACKWATER SYSTEM

This river system drains the southern limb of the London Basin westwards from Aldershot; the Wey system lies to the east, and the tributaries of the Kennet to the west. The Rivers Loddon, Whitewater, Hart and Blackwater are thus tributary to the longitudinal consequent Kennet–Thames. The Blackwater rather than the Loddon is the key stream of this group, because of its earlier extensions southwards towards Hindhead, from whence was derived the Lower Greensand chert pebbles, which can be traced in the terrace gravels all the way to the Thames terraces above Shiplake. From the studies of Bury and Linton it seems certain that these rivers originated on the post-Diestian marine plain, as transverse consequent streams which have become subject to the earlier mid-Tertiary structures in such a way that only in certain reaches can the original consequent lines of drainage be discerned (Fig. 2).

(a) High-Level Features

The highest and some of the most extensive gravel spreads of this region, below the Pliocene outlier of the Upper Hale plateau, occur in an arc from Crowthorne (845640) to Aldershot, forming the watershed between the Blackwater and the tributaries of the Wey. This is an area of military training grounds and it has not been possible to extend the detailed survey into the region, and the terraces have not therefore been mapped. It is clear, however, that these gravels display certain anomalous features. The highest altitude of 428 feet O.D. is found at Surrey Hill (888641) and from here the gravels fall in altitude without any perceptible break of slope towards Crowthorne in the west and southwards to Chobham Ridge, where they decline to 375 feet at Windmill Hill (917570). Across the gap at Deepcut (901568) the Fox Hills plateau rises southwards from *c.* 370 feet to nearly 390 feet at Surprise Hill (904516). From the physiography of these features it is difficult to escape the conclusion of the Geological Survey (1929), that there has been some post-Pliocene tectonic movement in this area, which has taken place subsequent to the deposition of the gravel, and aligned with several minor structures of mid-Tertiary age in the district around Deepcut. For this reason even that part of the plateau the slope of which conforms to the probable direction of drainage cannot justifiably be

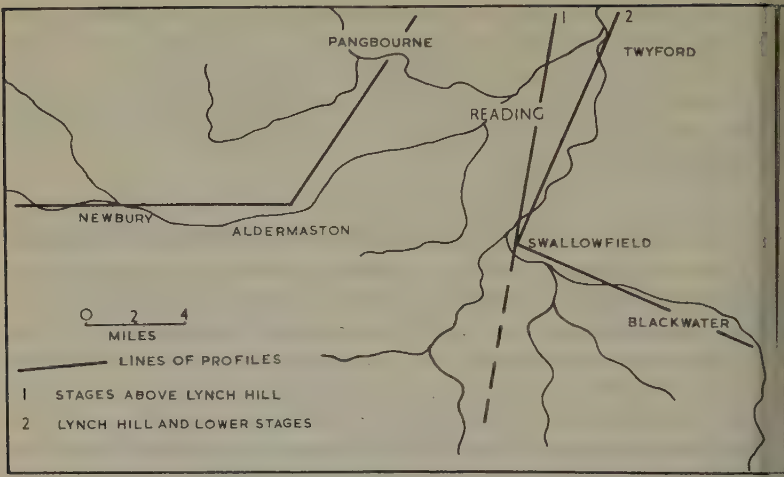

Fig. 4.

correlated with the undisturbed terraces farther west on the basis of altitudinal criteria. It is probable that the tilting suggested by the attitude of these deposits was influential in bringing about the deviation from the original consequent line now so evident in the middle Blackwater which is apparently guided by local flexures with a NW.–SE. strike (Fig. 4). In the following account of the principal terraces of the Blackwater system, attention is therefore confined to the middle and lower reaches of the river.

(b) The Winter Hill Terraces

The broad plateau of the Hartford Bridge Flats forms the interfluve between the Blackwater and Hart Rivers, falling with a gradient of eight feet per mile, in a direction a little north of west, parallel with the present stream, from 333 feet O.D. in the east to 306 feet O.D. in the west and 155 feet above the flood-plain. The magnitude and altitude of the plateau have led to direct correlations with the Silchester terrace of the Kennet which it resembles (White, 1909).

Upstream the terrace is possibly represented by the flat at Crawley Hill (884603) and may be continued by the Fox Hills to the south, but the likelihood of structural disturbance renders correlation inadvisable. Since it is known from the absence of chert in the Caversham–Henley, Winter Hill, channel that the Blackwater joined the Thames in the region of Henley, even at this early stage a pronounced swing in the course of the river must have been evident, if less abrupt, than at present. In this case the flat near Bearwood Lake (775685), with a height of 279 feet O.D., may represent a continuation of this terrace. It is also possible that the well-preserved flat at Farley Hill (751648) (three feet O.D.) may also be a part of this terrace. If this be so then a rapid slackening of gradient must have occurred in the neighbourhood of Farley Hill. Similarly, if this flat is a part of the Hartford Bridge Terrace then the Finchampstead Ridges, fifty feet higher and only three miles distant, must represent a higher stage; conversely the Farley Hills gravels may be of a later date.

The gravels of the Hartford Bridge Terrace form part of Prestwich's Southern Drift and are distinguished from those of the Kennet terraces by the presence of chert and cherty sandstone from the Hythe Beds of the Lower Greensand from the Weald. These gravels are generally more sandy than those of the Kennet. This mixture of finer grades with the flint and chert fragments appears to have contributed to the spectacular effects of frost-heaving found in the upper layers of these gravels, which are frequently contorted, with wedges and masses of sand occurring within the coarser material. These features are well displayed in a pit on Eversley Common (789595).

The correlation of the Silchester terrace of the Kennet with the Winter Hill of the Thames has long been maintained and is substantiated by the evidence adduced above (p. 422); whereas that between the Hartford Bridge stage of the Blackwater and the Winter Hill terrace is less well supported by positive evidence. The circumstantial evidence suggesting correlation with the Silchester terrace of the Kennet is, however, considerable; the similarity is striking both in the great extent of the plateaus making up the terraces and in the altitudinal relationships with the present streams. Close links between the Blackwater terrace and the Winter Hill Thames have not survived. It is apparent that the Hartford Bridge Plateau is continued downstream by the Bearwood flat and the change of gradient to about three feet per mile implied by this would bring the terrace into the Winter Hill Thames near Henley at 258 feet O.D. This is quite probable for the lower reaches of the river were, then as now, flowing against the structural trends of the region.

The possibility that the Upper Winter Hill terrace may be present in the Blackwater Valley depends upon the interpretation of the gravels capping the Finchampstead Ridges which form the divide between the Blackwater and northern streams, standing at 333 feet O.D., facing the Hartford Bridge Flats nearly three miles to the south. The characteristics and height of the Ridges are such as to suggest correlation with the larger plateau but lack of knowledge of the stream directions at this stage makes such a correlation precarious. From the most probable direction of flow, towards the WNW., orthographic projection across the valley indicates that the ridges stand some fifteen feet higher than the Hartford Bridge Flats and might therefore be related to a slightly higher stage (see above) comparable with similar features found in the Kennet and Thames Valleys (see p. 420).

(c) The Black Park Terrace

The possibility of the Black Park terrace of the Thames surviving upstream into the Blackwater Valley should be considered although Dr. and Mrs. Sealy thought that its upstream gradient from Slough would bring it into convergence with the Winter Hill terrace somewhere between Henley and the Thames Valley (1956). The gravels forming the high ground in the Shinfield Road area of Reading have a height of 267 feet O.D. and contain materials of the southern drift but no northern derivatives. They would seem to be too low to be correlated with the Winter Hill in the Blackwater Valley which across the valley at Bearwood is ten feet higher. Such a correlation would also necessitate that the Blackwater was at this time flowing at a lower level than the Thames at this point. If, therefore, this is part of the Black Park terrace its upstream course must remain in doubt. The gravels at Farley Hill are possibly a remnant of such a terrace having an upstream gradient of four to five feet per mile; on the other hand the latter flat may equally represent a continuation of the Hartford Bridge Flats (see p. 426). It is difficult to resolve this problem without further evidence from the gravels which do not show any sections.

(d) The Boyn Hill Terrace

Lying between the Loddon and Whitewater Rivers and south of the Blackwater is Heckfield Heath (720615), a well-preserved and extensive flat with a maximum height around 270 feet sloping northwards. This and other interfluve plateaus farther south form an important stage in the development of the drainage system, later than that represented by the Winter Hill gravels of the Hartford Bridge Flats. In the Blackwater Valley the Heckfield plateau grades upstream into an ill-defined valley bench at 279 feet O.D. on the flanks of Eversley Upper Common (767605), indicating a gradient of almost seven feet per mile. Southwards, into the country drained by the Loddon, Hart and Blackwater Rivers, this same terrace is continued at Hazeley Heath (755580), where the gravel surface falls northwestwards at a rate of twelve feet per mile from 291 feet O.D. to 278 feet O.D., and farther west a series of flats north and south of Rotherwick (712563) have a gradient of nearly fifteen feet per mile falling from 309 feet O.D. to 279 feet O.D. These gradients suggest that the gravels at Hook Common (716535) and Shapley Heath (756546) probably belong to the same stage, though their relation to former courses of the present streams must remain in doubt.

Downstream from Heckfield Heath little remains of this once extensive terrace, although small flats at 257 feet O.D. near Arborfield Court (758665), and above Bearwood Lake (774686), at 253 feet O.D., suggest its continuation downstream with a gradient of around four feet per mile, naintaining it at a height of 110 to 120 feet above the present stream. This uggests that this terrace grades into the Boyne Hill of the Thames in the egion of Sonning (757755).

The gravels on Heckfield Heath are about six feet in depth, and well tratified. Only one small section can now be clearly seen (718617) but it hows festooning in seams of sand and clay in four to six feet of gravel, eminiscent of the gravels at higher levels.

(e) The Lynch Hill Terrace

Below the Heckfield terrace, and some ninety feet above the present lood-plain, occur a series of scattered flats the largest of which are to be ound levelling the watershed between the Loddon–Blackwater basin and he valley of the Foudry Brook, just west of Spencers Wood (715664), at 24 to 228 feet O.D. Upstream into the Blackwater Valley, flats with a imilar relationship to the present streams occur on the watershed with the Iart River just south of Bramshill (753600) at 240 to 253 feet O.D. The eposits here can be examined in a large pit which reveal them to be preominantly sandy. In places five feet of sand, often current bedded, overlie wo to three feet of gravel. Gravel also occurs in the upper part of the eposit exhibiting similar contortions to those seen in the higher terraces.)ownstream, flats at Arborfield Cross (763672) and north of Shinfield 735683) suggest that these remnants are all related to a single stage in the evelopment of the river system, about ninety feet above the present treams and falling towards the Thames with a gradient of five feet per nile. This brings the terrace into the Lynch Hill terrace of the Thames ear Earley (745721) at 212 feet O.D. These lower terraces have a steeper lope to the Thames in the lower reaches resulting from the shortened ourse to the Thames, diverted from its Winter Hill channel between 'aversham and Henley to its more southerly route occupying the former ower Blackwater Valley.

(f) The Taplow Terraces

In common with the Kennet and the Thames there are two low terraces n the Blackwater Valley at about sixty and thirty feet above the flood-lain. These are well preserved in the Blackwater Valley, in places forming rue paired terraces—'paired sequence' contrasting with the 'longitudinal equence' resulting from postponed junction and with the 'parallel equence' resulting from uniclinal shift. This manner of preservation sugests a diminution in the volume of the river, probably following the second tage in the capture of the Blackwater headstreams, by the Wey.

These low terraces are preserved sporadically, mainly along the southern ank, from Blackwater (855597) to Bramshill. North of Eversley (775625) here is a fairly extensive flat at 213 feet O.D. displaying five feet of well-stratified sandy gravel. As the river turns northwards the terraces are preserved on the western flank of the valley at Riseley Common (727628) where they are at 210 feet O.D. and 180 feet O.D. respectively, and ther extensively along the eastern side from Swallowfield to Winnersh where they have declined to 180 feet O.D. and 160 feet O.D., and are clearly continued into the Upper and Lower Taplow terraces around Woodley (775736).

(g) The Flood-Plain

Although it is probable that a flood-plain terrace exists in the Blackwater and Loddon Valleys it is scarcely a traceable feature and has not been mapped. The senility of the present streams is very marked, exhibiting braided channels and 'broadwaters' which are partly artificial. Such features are a fundamental expression of the difficulty encountered by these rivers flowing against the dip to reach the Thames ten miles to the north.

These conclusions, regarding the correlation of the terraces described in the Kennet and Loddon–Blackwater Valleys with the terraces of the Middle Thames, are summarised in the accompanying Table I.

TABLE I. **Heights of Terraces above the present Flood-plains at selecte points**

	River Thames			River Kennet		River Loddon–Blackwater	
Terrace	Slough[1]	Medmenham[2]	Reading[3]	Aldermaston	Newbury	Shinfield	Eversley
Harefield	250	240	230	230	221	—	—
Rassler	223	210	205	—	200	—	—
Upper Winter Hill	200	184	175	172	165	—	—
Lower Winter Hill	189	162	153	155	155	154	157
Black Park	150	147	139	—	—	139?	—
Boyn Hill	115	120	108	—	—	123	120
Lynch Hill	70	83	90	—	—	90	94
Upper Taplow	40	44	65	—	56	54	56
Lower Taplow	25	—	30	37	30	31	38
Flood-plain	5	7	7	7	7	—	6

[1] After F. K. Hare. [2] After K. R. and C. E. Sealy.
[3] Mainly after K. R. and C. E. Sealy.

4. THE DEVELOPMENT OF THE DRAINAGE PATTERN

The earliest lines of drainage of the current cycle were superimposed from a thin Pliocene cover and were consequent upon the dip and pitch of the Pliocene trough. The Kennet–Thames formed the axial longitudina tream and the southern tributaries draining from the Hindhead massif of he Wealden area extended across the former Diestian plain to the Thames n the north (Fig. 5). The Crondall and Aldershot gaps testify to two treams draining from the Wealden region before capture by the subseuent tributaries of the Wey. These were probably the Hart and Blackwater espectively. The Whitewater may once have crossed the chalk through the olden Pot gap, but, with the Loddon, is essentially a stream draining rom the foot of the chalk escarpment. The Kennet and the Blackwater hich survived capture until a comparatively recent stage are thus the aster streams.

The subsequent evolution of drainage patterns in the area has been in esponse the three major influences, the increasing effects of the exhumed id-Tertiary structures resulting in uniclinal shifting, directional changes nd deferred confluence at stream junctions, the effects of successive

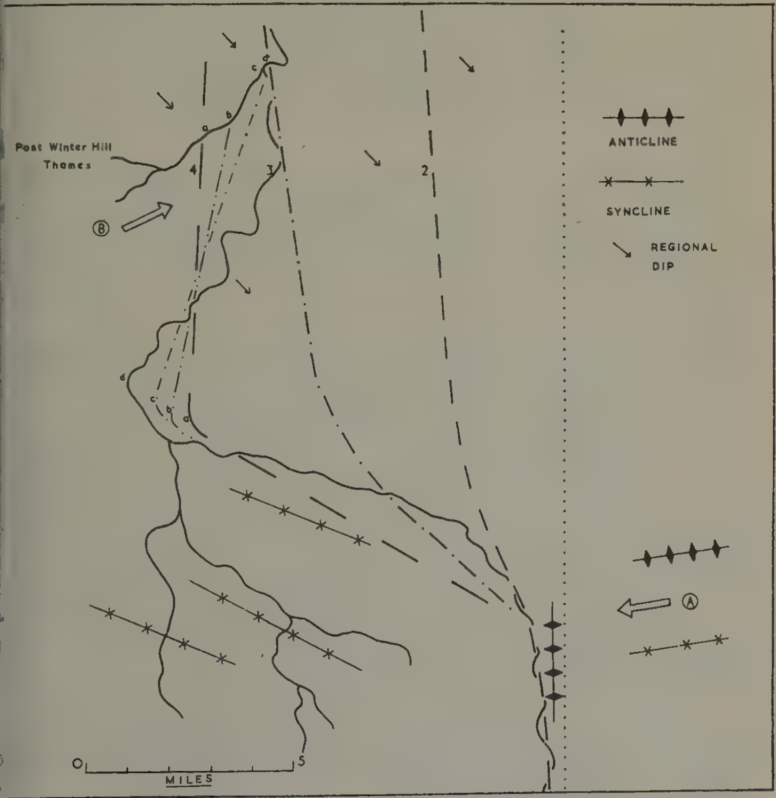

Fig. 5. The evolution of the middle and lower Blackwater

captures of the Wealden headwaters of the Hart and Blackwater by the Wey, and the results of the diversion of the River Thames itself.

The early stages of development of the Kennet appear to have been marked by uniclinal shifting as in the Thames. The absence of high terraces in the southern tributary valleys together with the doubtful character of the highest features around Crowthorne make it difficult to ascertain the manner of evolution of these rivers. It is clear, however, that by Winter Hill times the Rivers Hart and Blackwater were following the trend of minor folds influencing the direction of the streams in their middle reaches (Fig. 4). Whether former northerly courses have by stages deviated to the present almost east–west direction or whether the change was effected by capture it is impossible to tell. In either case the Winter Hill courses showed a considerable degree of adjustment to structure, a form of adjustment, however, which once the guiding structures died out westwards left the lower part of the Blackwater to make its way northwards to the Thames against the dip of the northern limb of the London Basin. These facts are demonstrated by the continued stability of the middle course in contrast to the repeated postponements of its junction with the Thames which have brought about the present elbow in the river near Swallowfield (Fig. 4).

The uniclinal shifting of the River Kennet resulted in the development and extension southwards of the Lambourn and the Pang, the former becoming adjusted to minor structures by repeated postponement of junction with the Kennet, while the Pang may have been diverted by capture to the west–east direction of the middle course, which is evident by Winter Hill times (Fig. 5 (2)).

The broad featureless plateaus of the Winter Hill terraces mark a period of prolonged planation throughout this region and their disposition and the character of the gravels make it difficult to be certain concerning the pattern of drainage at this time. The consolidation of the braided streams and subsequent downcutting, however, established a pattern of drainage which has changed little over much of the region to the present day. The course of the River Loddon at this time is problematical since no certain remnants of its Winter Hill plain have survived. It is possible that this river contributed to the great spreads of gravel present on the Silchester plateau but whether the channels of the two rivers were confluent or merely adjacent is difficult to tell. In the latter case the Loddon presumably followed a course northwards to enter the Thames channel near Caversham; this might account for the flats at Emmer Green, which stand at a height of 271 feet O.D. (722767). If the Loddon joined the Kennet it must have been diverted at a later stage, to form the forerunner of the Coley gap, through which the Kennet was later diverted.

Cognate to these problems is that of the diversion of the Thames itself om its Winter Hill channel to its present course via Sonning. This prob- m was considered by Mr. and Mrs. Sealy (1956) who, following Osborne hite, suggest that the Thames during the Winter Hill aggradation over- pilled into the adjacent Blackwater Valley, and that this course south of infield Heath was confirmed during the ensuing Black Park rejuvenation hen the former channel was finally abandoned. The second phase of this eory requires that while the Black Park rejuvenation progressed ten iles, as far as Reading, in the Tertiaries of the Blackwater Valley it made o headway at all in the Chalk of the Thames Valley. Even allowing for robable differential rates of erosion this seems unlikely. The following lternative explanation is suggested to comply with the available evidence. he gradient of the Winter Hill Thames below the Goring gap was as low two feet per mile, while in the Kennet and Blackwater the surface of this rrace quickly steepens to a figure of nearly ten feet per mile. The greater ope of the southern tributaries implies a higher ratio of load to volume, robably resulting from a copious supply of solifluxion debris from the halk downs, but only a moderate volume of water from the summer melts f local snow. The Thames, on the other hand, still fed by the meltwaters om the northern ice sheets, received both a large volume of water and a igh load, while its downstream course may have been impeded at this me by meltwater from the St. Albans ice lobe. Under these conditions its adient was much less. Further, amelioration of climate over the whole ea, resulting in the decay of the ice sheets, would have deprived the hames of its large supply of water as well as of its solifluxion waste. In the ibutary valleys, however, water from local snowfields may have been dequately replaced by increased precipitation resulting from the waning f the cold permanent anticyclone associated with the glacial climate. here is therefore the possibility of a period during which the tributary reams were able to degrade their Winter Hill channels while the Thames aintained its earlier level. Under such conditions the overspill suggested y White and others would have been confirmed before the knick point om the earlier glacial period reached the region of Henley.

With the post-Winter Hill erosion the separate identities of the tributary reams become apparent, and it is possible to define the limits of the later ood-plains which are frequently bounded by the bluffs leading down from e Winter Hill plateaus (Fig. 5 (3)). The Kennet continued to incise its bed to the loop around the Tilehurst plateau while the Blackwater con- stently and continuously postponed junction with the newly diverted hames, over the five miles between Reading and Wargrave from Black ark times until the present day.

Again it is the Loddon which cannot be charted with certainty; the roblem remains as to whether it continued to join the Kennet, as inferred y Hawkins for a later stage in its history, or whether it pursued an independent course to the Thames near Reading, as indicated by Dr. and Mrs Sealy (1956, fig. 5). There is little evidence, morphological or circumstantial, to substantiate either theory. The alignment of the middle Loddon with the Foudry Brook and the lower Kennet suggest a former course of that river, but the presence of Southern Drift in the Shinfield Road district of Reading argues against the existence of any stream other than the Blackwater entering the Thames in this area. It is therefore tentatively suggested that the Loddon joined the Kennet in the region of Silchester during Winter Hill times, continuously deferring confluence with that river until the junction lay between Burghfield (670685) and Theale (645711), along which line later entrenchment occurred. The process of postponement would clearly have been arrested by the existence of the higher ground around Tilehurst, but the tendency for the middle reaches of the river to swing farther towards Reading would have resulted from the structural controls. Remnants of a terrace, imperfectly preserved on the London Clay, around Trash Green (660691), at 215 feet O.D., some seventy feet above the River Kennet, may be evidence of this former course of the Loddon at the Lynch Hill stage (Fig. 1). The lip of the Coley gap, through which the Kennet now passes, is bordered by a remnant of this same terrace and it is suggested that towards the end of this stage a flood carried the Loddon via this route into the Thames, flowing at a level some ten to fifteen feet lower. Such a diversion would have lowered the Loddon Valley by rejuvenation, and this resulted at a later stage in the capture by or the diversion into the lower Loddon of the River Kennet.

This abrupt change in course of the lower Kennet has been described in detail by Professor Hawkins (1926) who thought that the explanation lay in the continued rise of the northern limb of the syncline. Yet structural factors seldom work alone, and the actual mechanism of the diversion was probably a form of capture or a result of the choking of the channel of the dwindling post-glacial Kennet by gravels deposited during the last glacial phase in such a way that the stream sought an easier route into the Loddon. By whatever mechanism, the diversion followed the formation of the flood-plain terrace, and the constriction caused by the Coley gap was probably responsible for the meandering in front of that feature of both the Kennet and the Loddon, which has given rise to the present wide flood-plain south of Reading.

The last event in the evolution of the existing drainage pattern appears to have been the capture of the Loddon by a pirate tributary of the Blackwater through the gap west of Swallowfield, leaving a dry gap, only a few feet above the present stream, north of Beech Hill (698644).

If correctly deduced this account presents a consistent tendency for the diversion of the rivers around Reading into more easterly channels (Fig. 6), following periods during which middle courses of the streams by uniclinal

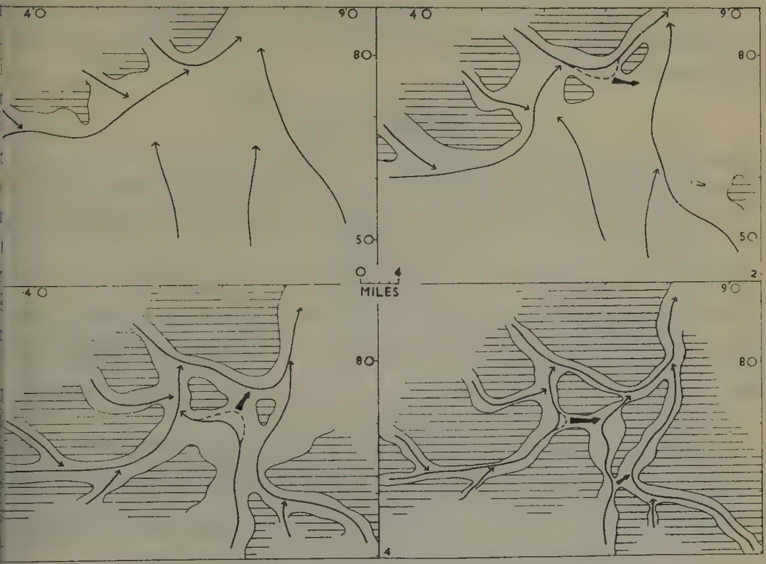

Fig. 6. The pattern of drainage evolution in the Reading area

hift and simple adjustment to structure brought the lower reaches into ncreasing conflict with the structural trends of the region: a picture of radually increasing maladjustment, mitigated in the Blackwater by a ontinued process of postponement of junction with the Thames, and ollowed in the Kennet and Loddon by sudden diversion. The structural nfluences may not have been static. Professor Hawkins has pointed out hat, 'the proved south easterly subsidence of the whole district . . . must rag the sensitive senile drainage with it'. If this assertion is correct then here is here an added factor to help explain the adjustments described n this study. But the structural trends themselves appear sufficient to nfluence the pattern of stream migration in this region, and the conditions f the streams during cold phases of the Pleistocene period were such that he changes in response to these controls could take place. Under the permanence of the stream courses in temperate climatic conditions such iversions appear strangely catastrophic, yet there is no evidence in this egion of direct intervention by the northern ice sheets as in the Thames alley.

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Notes on the Gault near Sevenoaks, Kent

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BSTRACT: Stratigraphical and faunal details of a temporary section through Upper ault (Upper Albian) at Kemsing, Sevenoaks, Kent, and of a borehole and extended xcavations in Gault at Greatness Lane, Sevenoaks, are given. Note is also made of an nteresting, though minor, tectonic feature exposed at Greatness Lane, and of the ariations in thickness of the standard Gault divisions across the Weald.

1. INTRODUCTION

URING 1959 and early 1960, roadside trenches, six to eight feet deep, and unning at right angles to the strike of the beds, were dug in Childsbridge ane, Kemsing, from the junction with Dyne's Road and West End, to hild's Bridge, a distance of approximately 800 yards. These trenches rovided some excellent exposures of the Upper Gault and the Gault/ halk junction, rarely seen in this district.

Whilst the Childsbridge Lane trenches were being excavated, the evenoaks Brick Co.'s Research Department put down a boring immediately o the north of the company's Gault pit at Greatness Lane which passed hrough practically the entire Lower Gault. Subsequently the company ommenced excavations on the northern side of the pit and has also extended the working a little to the west, exposing beds both higher and ower than those described by the writer previously (Milbourne, 1955).

The additional details of the Gault at Greatness Lane given here, when dded to the section published in 1955 and combined with the details of he Gault at Childsbridge Lane (half a mile to the east), give an almost omplete section of the Gault in this area.

2. THE CHILDSBRIDGE LANE EXCAVATIONS

Gault was not exposed along the entire length of the trenches, as the rift, which consists of large flints in a red clay matrix, was in places more than six feet thick, i.e. more than the depth of the trench. The following succession was observed:

Bed		ft.	in
	Chalk		
	Obscured by drift	4	0
6.	Bright green marl, sandy in texture due to abundant glauconite grains, and with nests of phosphatic nodules	6	0
5.	Light grey, highly calcareous marl	*c.* 10	0
	Obscured by drift	6	0
4.	Light grey, iron-stained marl	6	0
3.	Black phosphatic nodules, mostly fragmentary ammonites, in a glauconitic clay matrix	1	6
2.	Light grey, unfossiliferous clay	6	0
	Obscured by drift	30	0
1.	Light, blue-grey clay with limonitic fossils	5	0
	Obscured by drift	15	0
	Total	89	6

Beds 1 and 3 yielded abundant ammonites, listed below, but apart from a fragment of *Inoceramus* from bed 6, no fossils were obtained from the rest of the section.

Bed 3. *Callihoplites auritus* (J. Sowerby), *C. tetragonoides* Spath, *C.* spp. indet., *Epihoplites gibbosus* Spath, *Pleurohoplites* sp., *Euhoplites* cf. *boloniensis* Spath, *Eu. alphalautus* Spath, *Mortoniceras* (*Mortoniceras*) *inflatum* (J. Sowerby), *Mortoniceras* spp., *Cantabrigites cantabrigensis* Spath, *C. minor* Spath, var. *serpentina* Spath, *C.* aff. *subsimplex* Spath, *Prohysteroceras* (*Goodhallites*) *goodhalli* (J. Sowerby), *Hysteroceras carinatum* Spath, *H. subbinum* Spath, *H. orbignyi* (Spath) late mutation, *H. bucklandi* (Spath), *H.* spp., *Hengestites aplanata* Casey.

Bed 1. *Hysteroceras varicosum* (J. de C. Sowerby), *H. binodosum* (Stieler), *H. binum* (J. Sowerby), *Euhoplites alphalautus*, *Epihoplites* sp. juv., *Idiohamites favrinus* (Pictet).

It is clear that bed 1 is in the *varicosum* Subzone of the *Mortoniceras inflatum* Zone, i.e. equivalent to some part of Bed X of the Folkestone Gault, whilst bed 3 is a condensed representative of the *auritus* and *aequatorialis* Subzones of the *inflatum* Zone, i.e. beds XI and XII of the Folkestone Gault, similar to the nodule bed at Horish Wood, on the Maidstone by-pass, described by Casey (1959), and now known to occur also at Aylesford and Wrotham, Kent.

At the three latter localities the *auritus-aequatorialis* Subzone nodule bed is directly underlain by blue-grey clay containing pyritic and limonitic ammonites of the late *varicosum* Subzone. At Childsbridge Lane, however, the six feet of clay seen beneath the nodule bed was light grey, marly and unfossiliferous.

It was not possible to demonstrate the thickness of the *varicosum* ubzone at Childsbridge Lane owing to the drift obscuring part of the ection, but blue-grey clay yielding the typical *varicosum* fauna occurs hirty-six feet below the *auritus-aequatorialis* nodule bed, and since the *aricosum* Subzone is only five feet thick at Folkestone, it would appear hat there is a considerable thickening at Childsbridge Lane. The only ntermediate exposure where the full thickness of the Subzone can be lemonstrated is at Wrotham, five and a half miles to the east of Childsridge Lane, where it is twenty-five feet thick.

I would hesitate in assigning the unfossiliferous clay beneath the nodule ed at Childsbridge Lane to the *varicosum* Subzone, however, since the ighly fossiliferous nature and uniform lithology displayed by the Subzone t other localities in the northern and eastern parts of the Weald is evidence gainst such a correlation. It is possible that the unfossiliferous clay of bed at Childsbridge Lane belongs to the *auritus* Subzone, the *auritus-equatorialis* Subzone nodule bed incorporating part only of the Subzone.

On the basis of physical properties, bed 6 of the Childsbridge Lane ection is tentatively correlated with the Chloritic marl of the Cenomanian tage. None of the Gault seen at Kemsing was in the Upper Greensand acies.

3. GREATNESS LANE

(a) Lower Gault

The boring referred to above began in the upper *lautus* Zone of the ower Gault, i.e. bed 17 of Milbourne (1955), and finished in the lower *entatus* Zone, ten feet below the nodule bed at the top of the *dentatus-pathi* Subzone, i.e. bed 2 of Milbourne.

Three inches above the bottom of the borehole, i.e. nine feet nine inches elow bed 2, the core showed another thick layer of phosphatic nodules rom which several fragments of *Hoplites*, including *H. rudis* Parona and onarelli and *H. persulcatus* Spath, were obtained. From this it is clear hat the *dentatus-spathi* Subzone is at least ten feet thick at Greatness Lane. oring unfortunately stopped at this level and it is not, therefore, possible o say whether there is any representative of the *benettianus* Subzone resent in the area. The cores from this borehole are preserved at the ffices of the Sevenoaks Brick Co.

In July 1957 excavations in the eastern end of the pit exposed the top six eet of the *dentatus-spathi* Subzone and from this exposure the author and Mr. R. Beaney collected many fossils and drew up the following section:

Bed	ft.	in.
2. (Milbourne, 1955)		4
1. (pars Milbourne, 1955)		4
K. Crushed shells and sparse phosphatic nodules		1
J. Dark grey clay	1	0

Bed	ft.	in.
I. Seam of large, crushed *Inoceramus* with soft, brown phosphatic nodules		4
H. Grey clay with crushed fossils		10
G. Seam of crushed shells and phosphatic nodules		2
F. Grey clay with crushed shells		8
E. Crushed shells and sparse phosphatic nodules		2
D. Buff clay		7
C. Crushed shells and sparse phosphatic nodules		1
B. (pars). Grey, ironstained clay	1	5
SEEN IN CORES FROM BOREHOLE		
B. (pars). Grey, ironstained clay, a further	3	9
A. Black phosphatic nodules, mainly fragments of *Hoplites* ...		2
Grey clay, seen for		3
Total	10	4

Hoplites occurred in all the nodule and shell beds and the preservation of these and other fossils was remarkably good, all specimens retaining the test. An example of the Nautiloid *Eutrephoceras* sp., from bed I, exhibited the original colour banding of the shell. The colour bands, which were similar in colour and contour to those of the recent *Nautilus pompilius* Linn., faded rapidly on exposure to the air and sunlight.

Bed G is remarkable for the number of *Protanisoceras* which it yields, twenty-two specimens having been found in a single afternoon.

Further collecting in bed 2 at Greatness Lane has yielded many ammonites additional to those listed by me in 1955 (239), notably *Mojsisovicsia delaruei* (d'Orbigny) and *Oxytropidoceras* aff. *multifidum* (Steinmann). These and other 'exotic' forms, hitherto unknown from the English Gault, have been obtained by the writer from an equivalent horizon at Wrotham, Kent, and have been listed by Casey (1959, 206, 207).

The writer does not agree with the raising of the upper limit of the *dentatus-spathi* Subzone to a position within the *intermedius* Subzone, as suggested by Owen (1958, 161). The thick layer of phosphatic nodules (bed 2) occurring at what has always been considered the summit of the *dentatus-spathi* Subzone (e.g. Wright, 1946; Casey, 1954; Milbourne, 1955), provides an easily recognisable and well-defined limit which extends at least around the northern part of the Weald. Furthermore, neither *Hoplites dentatus* or *H. spathi* occur above this nodule bed and this alone would seem to forbid the reference of any overlying strata to the *dentatus-spathi* Subzone.

The exposures of the *dentatus-spathi* Subzone are now (1960) beneath water-level in the flooded part of the pit.

(b) Upper Gault

The highest beds of the Gault to be seen at Greatness Lane are exposed in the extreme western end of the pit where they are brought down by the udden increase in the westerly dip of the strata, noted previously (Mil-ourne, 1955, 236). Extension of the workings to the westward has now xposed thirteen and a half feet of clay referable to the *orbignyi* Subzone. etails are as follows:

Bed		ft.	in.
46.	Fawn-grey, unfossiliferous clay	1	6
45.	Fawn-grey clay with many crushed *Actinoceramus sulcatus* (Parkinson) and scattered phosphatic nodules	2	0
44.	Soft, brown phosphatic nodules		1
43.	Fawn-grey clay with scattered nodules		6
42.	Crushed shells and sparse phosphatic nodules. Many large, crushed *Mortoniceras*		2
41.	Fawn-grey clay with phosphatised and limonitised fossils and with lenticles of clay-ironstone at the base	2	6
40.	Crushed shells and sparse phosphatic nodules		1
39.	Hard, marly clay with lenticles of clay-ironstone and calcite	1	0
38.	Crushed shells and small phosphatic nodules		1
37.	Fawn-grey clay	1	6
36.	Crushed shells and sparse phosphatic nodules		1
35.	Light, fawn-grey clay	1	2
31 to 34.	Described previously (Milbourne, 1955, 236) Total	3	4
	Total thickness of *orbignyi* Subzone exposed	13	6

The following ammonites were collected *in situ*:

Bed 45. *Hysteroceras orbignyi* (Spath), *H. subbinum* Spath, *Mortoniceras pricei* Spath, *Hamites intermedius* J. Sowerby, *H. intermedius distincta* Spath.

Bed 44. *Hysteroceras orbignyi*, *H.* aff. *varicosum* (J. de C. Sowerby), *H.* sp. nov., *Euhoplites inornatus* Spath, *Eu. subcrenatus* Spath, *Epihoplites trifidus* Spath.

Bed 42. *Hysteroceras orbignyi*, *H. subbinum*, *H.* aff. *choffati* Spath, *H.* sp. nov., *Epihoplites* aff. *compressus* Parona & Bonarelli, *Hamites intermedius*, *Mortoniceras pricei*, *M.* (*Deiradoceras*) *bipunctatum* Spath, *M.* (*Deiradoceras*) *albense* Spath, *M.* spp.

Bed 40. *Hysteroceras orbignyi*, *Euhoplites inornatum.*

Bed 38. *Hysteroceras orbignyi*, *H. subbinum*, *Euhoplites inornatum*, *Anahoplites planus* (Mantell), *Hamites intermedius. Mortoniceras pricei.*

Bed 36. *Hysteroceras orbignyi*, *Euhoplites inornatum*, *Eu.* sp. nov., *Epihoplites trifidus*, *Beudanticeras beudanti* (Brongniart).

Bed 34. (Additional to those listed in 1955 (237)). *Hysteroceras* (*?Dipoloceras*) *symmetricum* (J. de C. Sowerby), *Hysteroceras* spp. nov., *Mortoniceras pricei*, *Beudanticeras beudanti*, *Psilohamites bouchardianum* (d'Orbigny), *Psilohamites* ? sp. nov.

4. TECTONIC FEATURES

The occurrence of a sudden increase in the westward dip of the Gault in e extreme western end of the pit at Greatness Lane has already been ted. Extension of the workings has now revealed that, following the sudden change in dip, there is a series of three small faults with upthrows, on the western side, of two, one, and two feet respectively. These throws are clearly seen at the junction of the *intermedius* and *niobe* Subzones, where there is a sharp colour change in the clay.

A feature of interest associated with the faults is the apparent compression, followed by gradual re-expansion to normal, of the clays immediately west of the fault lines. This feature is well seen in the clays of the *lautus* Zone where it produces a small monoclinal effect to the west of each fault.

The most westerly of the faults deserves special mention since the throw at the *cristatum* Subzone nodule bed (bed 18) is no more than six inches, whereas the throw in the *dentatus* Zone is two feet. This difference in throw is seen to be due to the variable resistance that the clays at different levels have offered to the compression effect of the faulting, i.e. where the clay has offered most resistance the throw of the fault is greatest. In keeping with this it is seen that the ensuing monoclinal effect is greatest in the clays which have suffered most compression.

5. CONCLUSIONS

The combined sections at Greatness Lane and Childsbridge Lane show that the Gault to the north of Sevenoaks has a minimum thickness of 138 feet. Allowing for the break in the continuity of the sections (involving the summit of the *orbignyi* Subzone and the base of the *varicosum* Subzone) and the unknown thickness of clay between the *dentatus-spathi* Subzone and the *mammillatum* Zone at the top of the Folkestone Beds, it is probably much nearer 150 feet.

Casey notes (1954, 266) that twenty feet of clays of the *dentatus-spathi* Subzone occurred in the Metropolitan Water Board boring at Brasted, four miles west of Greatness Lane, and it appears probable, therefore, that the thickness is between ten and twenty feet in the Sevenoaks area as a whole.

Of this total of 150 feet, 50 feet belong to the Lower Gault and 100 feet to the Upper Gault. At Folkestone, the type locality for the English Gault, these divisions are twenty-eight and a half and seventy-seven and a quarter feet thick (Jukes-Browne, 1900). There is, thus, considerable thickening of both Lower and Upper Gault in the Sevenoaks area compared with the type locality.

The gradual westerly increase in the thickness of the Gault in the Weald has been noted by previous observers (Jukes-Browne, 1900, Spath, 1943), but it is only in recent years that work has commenced on a detailed study of the variations in thickness of the individual Subzones. At present it will suffice to note that the *dentatus-spathi* Subzone is very variable; the *intermedius* Subzone, *subdelaruei* horizon of the *niobe* Subzone, *orbignyi* nd *varicosum* Subzones increase rapidly westwards; whilst the *lautus-itidus*, *daviesi*, *auritus* and '*aequatorialis*' Subzones are attenuated. The *iobe* and *cristatum* Subzones maintain a fairly constant thickness throughut the entire area, but increase slightly in western Sussex, whilst the new ubzone within the *lautus* Zone, mentioned previously (Milbourne, 1955, 41), expands rapidly westwards.

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An Example of the Origin of Coal-Balls

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BSTRACT: Several localities in the Herrin (No. 6) Coal, worked in the Sahara pencast Mine, Harrisburg, Illinois, possessed accumulations of coal-balls. These xposures provided significant evidence relating to the general development of coal-alls, and the petrification of wood. As in almost every other recorded occurrence of oal-balls, these accumulations were associated with the development of marine eposits.

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1. GEOLOGICAL SUCCESSION

HE HERRIN (No. 6) Coal belongs to the Middle Pennsylvanian, and rests pon three feet of clays, which formed the seat-earth (see Fig. 1). Beneath hese clays occur some fourteen feet of permeable sandstones, which are onsidered to play an important role in the formation of the coal-balls. The reservation of Stigmarian rootlets, together with the well-banded nature f the basal parts of the coal suggests an *in situ* origin for the seam. ronounced bands of fusain saturated with marcasite characterise what is therwise a normal vitrain-clarain-fusain seam of coal. Such developments f marcasite are also associated with other authigenic minerals such as alcite and quartz. The upper surface of the Herrin Coal is fairly uniform nd characteristically concealed by dark calcareous shales containing ellets and shreds of vitrainous coal. This suggests a somewhat uniform ubmergence of the swamp beneath sea-level. When these argillaceous eposits had settled, limestones developed in the clearing waters. The base f these limestones is extremely irregular, often filling fissures and solution avities in the calcareous argillaceous deposits beneath. Moreover, these imestone deposits probably provided the source of secondary calcite which ills joints (cleats) in the Herrin Coal and also cracks in the coal-balls found n this seam.

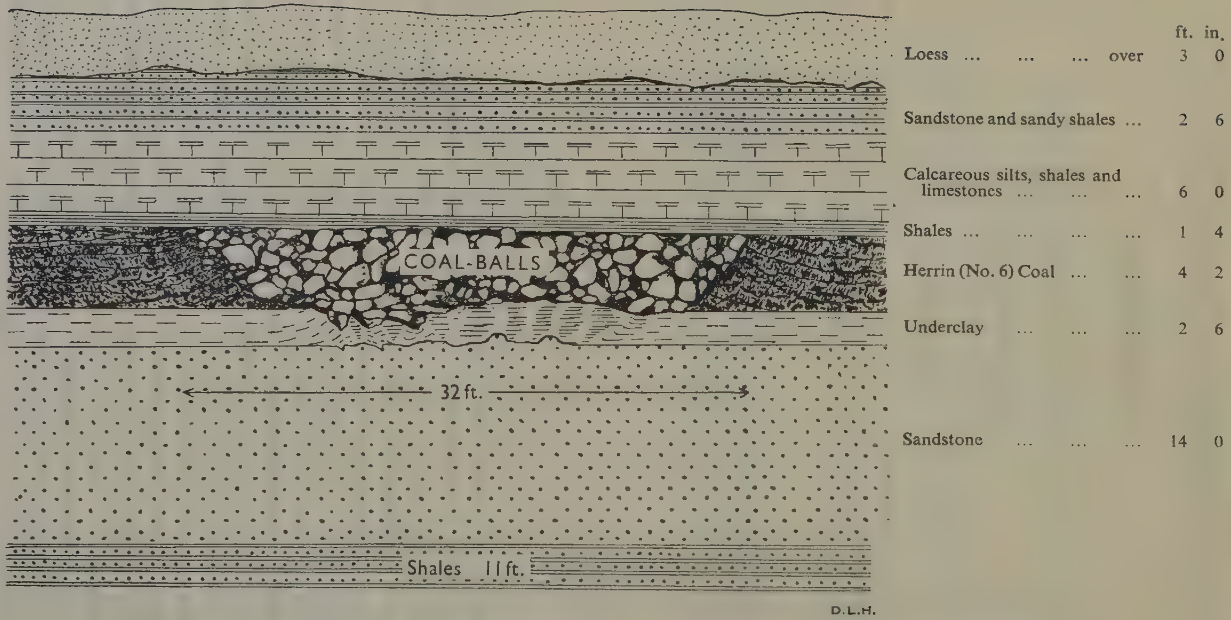

ft. in.
Loess ... ... over 3 0
Sandstone and sandy shales ... 2 6
Calcareous silts, shales and limestones 6 0
Shales 1 4
Herrin (No. 6) Coal 4 2
Underclay 2 6
Sandstone 14 0
COAL-BALLS
32 ft.
Shales 11 ft.
D.L.H.

Fig. 1. Section illustrating the relationship of coal-balls to the Herrin (No. 6) Coal and the associated strata

2. COAL-BALL LOCALITIES

A considerable number of coal-ball localities were encountered during e development of the extensive Sahara Opencast Pit. Some distance away, nderground in the Clarkson Mine, Cady (1936) recorded the presence of ockets' of coal-ball accumulations. Schopf (1936) also described them nd indicated that they consisted of poorly preserved logs of wood and alled-up, heterogeneous accumulations of plant material, frequently in an xcellent state of preservation. In addition, he mentioned that some of these oal-balls were encrusted with calcite (cf. Plate 19 A) and indicated that the me was deposited at two successive stages separated by an interval of coal ompaction.

It is difficult in underground workings to study the relationship between oal-balls and the coal seam itself. In the opencast workings of the Sahara it the plans indicated that a large number of hollows filled with workable oal-balls had been encountered. At the time of this investigation at least hree of these localities were clearly exposed in the workings, but others hich were partially concealed beneath debris supplied a good deal of onfirmatory data concerning their origin. It was also fortunate that the errin Coal had not been disturbed by structural factors, even though in ther parts of the pit there were many examples of small-scale folds and aults affecting the seam and associated deposits.

The largest locality exposed at the time of this investigation measured 71 feet across (see Fig. 2). In the central and uppermost parts of this ocality there was a slight depression, about forty-six feet in diameter, ccupied by a mixture of calcareous shales and coal debris. This was vidently a primary feature of the hollow which was filled with coal-balls rior to the submergence of the coal beneath sea-level. Moreover, it was oteworthy that around the hollow the coal-balls were piled up above the evel of the coal seam as a concentric convex bank (Fig. 2).

About 300 yards away another large locality, thirty-two feet in diameter, ad a dome-shaped upper surface formed by banks of coal-balls embedded n coal debris and sediment (Fig. 1 and Plate 19 B). Nearby, a smaller locality, nine and a half feet diameter at the surface, was clearly funnel shaped nd the result of a hollowed-out portion of the seam which had become illed with coal-balls, coal debris and sediment (Fig. 3). A considerable mount of silt and carbonaceous clay bonded the coal-balls in this locality, nd such sediments had also seeped into the adjacent joints in the coal. At a earby locality, a partially exposed hollow in the coal contained laminated eposits of carbonaceous siltstones and mudstones which were firmly elded to the sharply eroded sides of the Herrin Coal. Not more than six eet away from the coal these sediments surrounded *ad hoc* accumulations f coal-balls.

In all the localities exposed in these extensive opencast workings, the

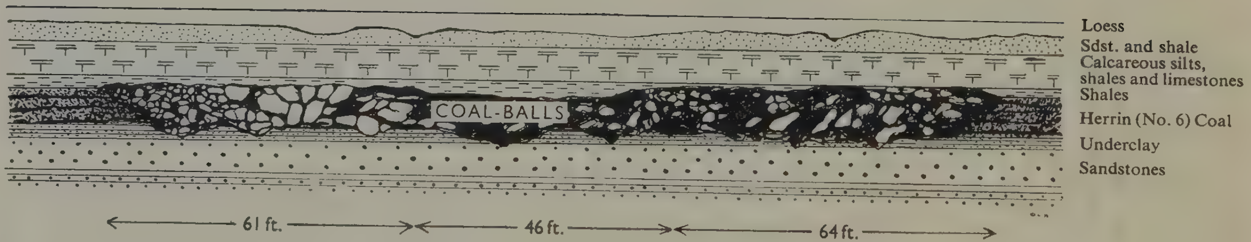


Fig. 2. Section showing an extensive area of coal-balls with the central depression occupied by shales

oal-balls were always restricted to somewhat circular hollows eroded in e Herrin Coal. Wherever it was possible to examine the base of these ollows they were cut through the seam. Moreover, the clays and under-ing sandstones were highly disturbed—evidently by springs developing rough the coalifying humus. Sufficient hydrostatic pressures to form low-wells were probably developed in the extensive deposit of sands over hich the Herrin Coal was developed.

Another feature common to all these coal-ball localities was their nvex surfaces, elevated above the top of the Herrin Coal. This was early a primary feature of the mode of accumulation of the coal-balls, nce the overlying shales progressively concealed these deposits. Moreover, e succeeding marine deposits showed no signs of disturbance which ight be connected with the presence of these hollows in the Herrin Coal lled with coal-balls. Within these hollows, the coal-balls ranged from mall pellets of *Stigmaria*, an inch or so in diameter, to large broken trunks f petrified *Calamites* several feet long, and all of them were encased in oppleritic coal. These rounded masses were all loosely cemented in a ixture of carbonaceous silts and clays together with a considerable mount of fragmented coal. These deposits were subsequently encrusted ith calcite, marcasite and gypsum, probably during the development of e overlying calcareous, estuarine and marine deposits. The haphazard istribution of the large blocks of petrified wood (one measured four and half feet long) does not suggest that they were derived from distant calities. On the one hand, it must be acknowledged that they resembled iles of driftwood which had accumulated in these circular hollows. Such a nclusion would be upheld if petrification was fulfilled after the hollows ad been constructed in the Herrin Coal. On the other hand, if such had een the case it would be reasonable to expect to find complete trunks, any feet long, among such large accumulations of petrified wood. In all

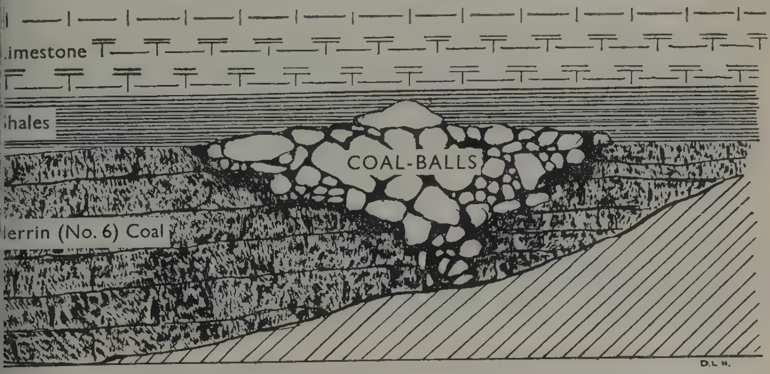


Fig. 3. Circular pipe-like hollow in the coal filled with coal-balls

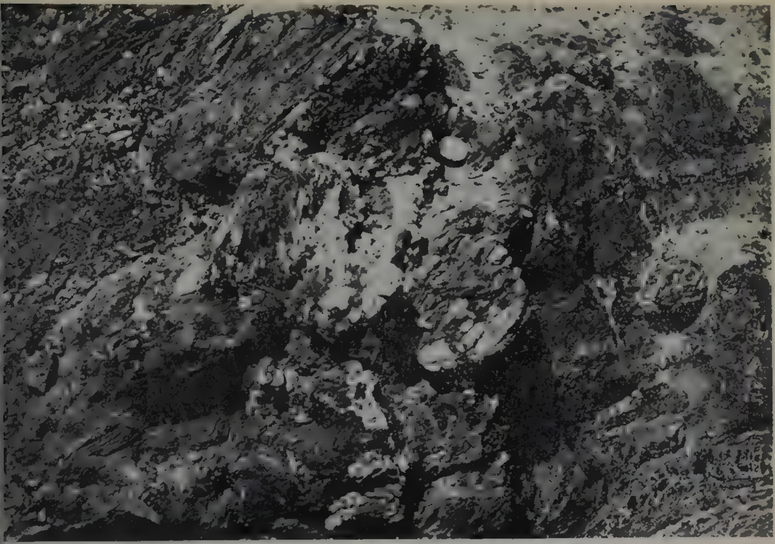
cases the trees had been severely fragmented after petrification and th internal fractures filled with doppleritic coal and secondary minerals, such as calcite. Supporting such a concept is the clear evidence of the existence of the hollow prior to the encasement of the petrified wood in dopplerite to form the coal-balls. On balance it would appear that the final stage of coal-ball formation marked the infilling of these circular hollows. The petrification of the roots and trees might well have been completed prior to the formation of the coal and the hollows in the seam. Under such condition localised areas of *Calamites* trees could have been petrified, probably by excessive quantities of lime produced by petrifying springs in the coal-forming swamps. Collapse of such trees would tend to dislodge and disintegrate the petrified roots and boles of these trees. Such collapse would be aided by the hydraulic forces involved, which would also cause these petrified masses to move slowly within the spring-loaded area and so become coated by the gelating humic exudates of the coal-forming swamp

3. PETROGRAPHIC DESCRIPTION OF THE COAL AND COAL-BALLS

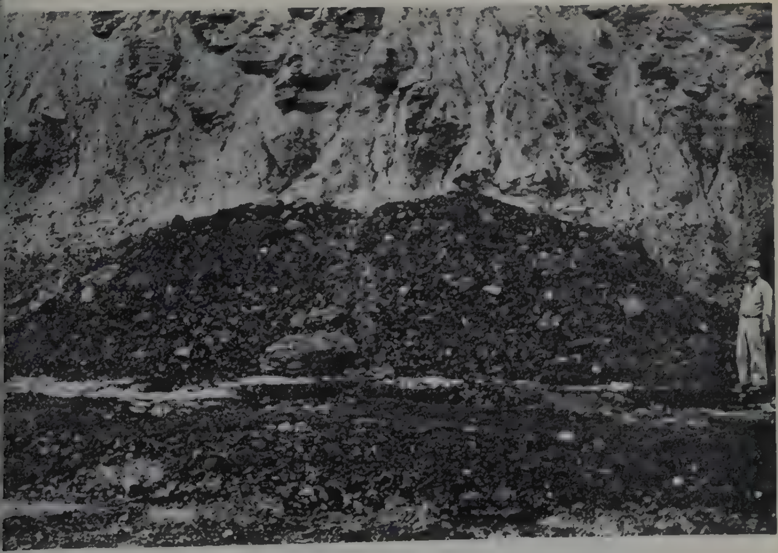
The Herrin (No. 6) Coal is a normal banded deposit of vitrain, clarai and fusain—the latter often highly saturated with marcasite, in bands ofte as much as one quarter of an inch thick. In addition, secondary veins o calcite and marcasite abound in the seam, but these are mainly associate with the joints (or cleat) in the coal. The marcasite bands developed i fusain are probably of authigenic origin, developed by the release o ferrous sulphides during the gelation of the humates to vitrain—the porous structure and the adsorptive character of the fusain bands providing the obvious stimulus for crystallisation from these solutions within the coal

Within the coal-balls themselves, the replacement of plant tissue was principally accomplished by calcite. This fills the cells, and also partly replaces the cell walls. Many of the cell walls survived calcification (Plate 20 A) and a chromatographic analysis of these indicated the presence o naphthenic humates. Of considerable interest are those cells replaced by marcasite. This mineral also sealed up some of the small fractures withi the calcified wood. Replacements of cellular tissue by marcasite appear to be adventitious and coincident with calcification, whilst the sulphides penetrating the cracks in the petrified wood might well be associated with the late-stage encrustations of marcasite on the surfaces of the coal-balls

Despite the fact that the cells were completely filled with rhombohedra calcite no obvious signs of cellular deformation were evident. This suggested a transfer of calcium carbonate through the cell walls to replace the break-down products of the nucleus. It is unlikely that this could have been achieved as a process of putrefaction of decaying plants, but suggests phenomena akin to those responsible for the formation *in situ* of petrified



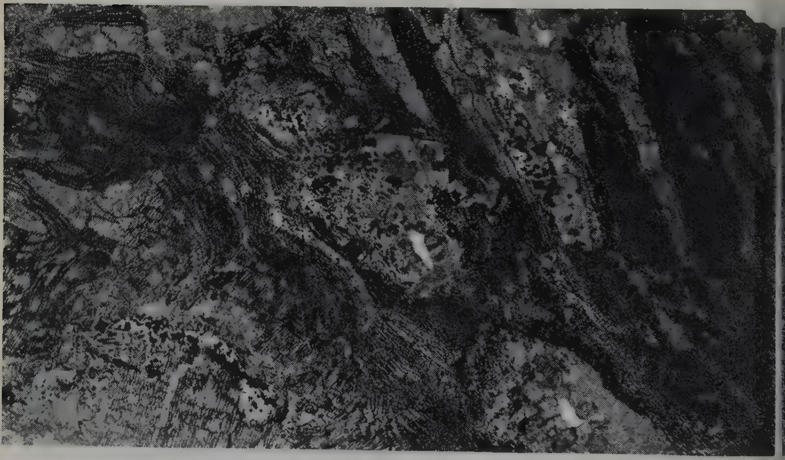
. Encrustations of marcasite and calcite on the outer surfaces of a group of large oal-balls



B. Convex upper surface of coal-ball accumulations. Note undisturbed nature of overlying calcareous shales and limestones and the size of individual coal-balls as indicated by the one dislodged into the foreground

[*To face p.* 450

A. Partially coalified tissue surrounded by doppleritic coal. Remainder of the cellular tissue almost completely calcified. Mag. × 10



B. Coal-ball tissue showing secondary veins of calcite and authigenic marcasite. Mag. × 10 ees. This is usually accomplished by the excessive intake of dissolved ineral matter—in this case calcium carbonate. Those parts of the tissue hich survived calcification passed through the normal stages of coalifica- on to form dopplerite and fusain (Plate 20 A). This is a common feature petrifying springs, especially those highly charged with carbonates and lphides in saline waters. When the tree finally collapsed the petrified ots and trunks would conceivably be upturned and fractured so that the gular fragments could then become encased in the colloidal humates rming in the swamp. In this way the internal cracks would also become led with these solidifying humates which are now found permeating the trified remains in the coal-balls. Contraction of this material during the al stages of gelation would produce secondary crack-systems in which lcite and marcasite would develop (Plate 20 B). During these latter ages the petrified roots and fragmented tree-trunks would be slowly tated by the surging springs and thus become coated with gelatinous udates from the surrounding coalifying deposits. In this way even the rge logs would become incorporated in doppleritic coal and all these unded objects piled up to yield the convex surfaces above the level of the al seam.

Analyses of the petrified *Calamites* yielded the following results:

	Specimen 1	Specimen 2
CaO	23.47%	23.37%
MgO	0.89%	1.06%
Fe_2O_3	5.25%	5.25%
P_2O_5	0.12%	0.16%

rom this it is evident that the high percentage of calcium and low agnesium would appear to rule out the possibility of the calcite origina- ng within the plant cells as a residuum of cellular decay. Both these salts ould have been present in approximately equal quantities if such had en the case. It is therefore necessary to postulate the entry of calcium to place the cell protein during petrification. No trace of crystalline calcium hosphate could be detected by X-ray powder photographs. This suggests at the phosphorus was derived from the breakdown of the cellular tissues the tree and remained behind as phosphate. This further enhances the ossibility of the intake of calcium through the cell walls to form calcite uring the degeneration of the proteins as an integral part of petrification.

4. DISCUSSION AND CONCLUSION

Despite the fact that coal-balls have been found in a variety of different ircumstances in coal seams or in the associated sediments, their principal haracteristics can be summarised as follows:

(a) Coal-balls are almost invariably found in coals, or derived from eams by subsequent erosion.

(b) They are more often than not associated with marine deposits an that such saline conditions were never far distant from the coal swamp during coalification.

(c) Coal-balls are often aligned along planes of stratification in a coal o occur as nests and ridges associated with disturbed portions of the sean

(d) Coal-balls are always surrounded by doppleritic coal or vitrain whic has congealed from a colloidal condition.

(e) Although of polished, rounded configuration they possess none o the properties of concretions.

(f) Mineralisation of plant fragments within coal-balls always appear to have been complete prior to encasement in doppleritic coal.

(g) They usually form centres for subsequent mineralisation, usually o calcite, pyrite, gypsum and quartz.

Of considerable importance is the stage at which the plant-remain become petrified and the way in which they escaped the more norma processes of coalification. Modern examples of petrifying plants are usuall associated with soils and subsoils wherein excessive quantities of dissolve mineral-matter are taken up by the roots and then progressively precipitate as a pathogenic replacement of the plant tissues. The exact mechanism o such petrifying processes is still imperfectly understood, but it is widel recognised as being associated with highly mineralised waters, an particularly with saliniferous swamps. This prompted the observations o Stopes & Watson (1908) who maintained that marine conditions wer conducive to such perfect preservation of the plant structures so commonl found in the petrified plant remains of coal-balls. It would therefore appea that the almost universal association of coal-ball formation with marin conditions has not been accidental. Innumerable examples have bee described from all the European Carboniferous Coalfields which hav emphasised how near sea-level were the horizons on which petrified wood developed to provide fragmented material for the formation of coal-balls Obviously such indurated masses would withstand a considerable amoun of transportation, and it is not surprising that they are found so frequentl in the sediments overlying the coal-forming horizon from whence they wer originally derived. The examples revealed in the Sahara Opencast Pi simply indicate one way in which such coal-balls have been preserved in hollows generated as blow-wells in the coal-forming swamp. On the othe hand, the contemporaneous processes of petrification were probabl identical with all other examples of coal-ball formation—a process stimulated by excessive mineralisation caused by spring waters and changes in the level of the water-table.

Feliciano (1924) investigated the coal-balls of the Illinois basin and concluded that they were formed *in situ* and that the plant structures were reserved from rapid decay by the peaty acids exuding from the sur-ounding coalified material. He points out that stigmarias, which are so ommon in coal-balls, indicate a static mode of origin, yet he appreciates at the sedimentary portion of the structures are now divorced from the eat-earths and the overlying sediments of the coal seams. Noe (1931) oted that the majority of coal-balls occur in the Pennsylvanian strata and articularly in the uppermost divisions. Moreover, they seem to be ssociated with those Middle Pennsylvanian deposits affected by sub-aerial rosion, suggesting that estuarine rather than stagnant swamp conditions revailed throughout these areas.

The coal-ball localities described from the Sahara Mine exhibit charac-eristics which emphasise that the petrified plants were developed during e 'peat-stage' but independent of the processes of coalification. Since all ypes of petrified plant material possessed fractured surfaces, they probably epresent the collapsed remains of trees and undergrowth subjected to xcessive secretion of mineral matter. The seepage of marine waters into andstone aquifers beneath the seat-earths would produce saline conditions hich would cause plant decay. Moreover, they would induce excessive oncentrations of salts capable of petrifying root and wood in restricted ocalities. Saline diffusion within such aquifers would also stimulate springs, o that trees and undergrowth might locally become mineralised to petrified ood. This could account for the extraordinary perfection of this fossilised lant-debris, and might account for the almost universal absence of fossil-sed leaves from coal-balls, which would naturally have disappeared long efore the break-up of the petrifying plants. Such hydrostatic forces would lso account for the upheaval of these petrified trees and plants so that ven the roots were upturned and fragmented. Finally, forces of this kind ould be required to account for the assembly of this petrified debris into estricted areas and at the same time enabling them to be encased in elatinous humic exudates from the coal swamp. Fast moving waters roding channels would inhibit such accumulations of dopplerite coal round such irregularly shaped fragments. Moreover, since these petrified emains show no signs of the erosion which could be expected from such an nvironment, it is unlikely that they were drifted as such into these circular epressions in the Herrin Coal. Finally, the somewhat circular basins ithin the Herrin Coal, wherein these coal-balls were accumulated, support e conclusion that these petrified plant remains were developed around ineralising springs or blow-wells. The irregular and inwardly sloping ides of these hollows and the funnel-shaped structures cut through the oal are typical of structures developed by such powerful springs. More-ver, the almost invariable occurrence of marine deposits suggests a orrelation of these structures with modern maritime blow-wells. The ands below the seat-earth of the Herrin Coal probably provided an excellent aquifer for the development of the interplay between the hydro static forces of saline and estuarine waters beneath the coal-forming swamp In this way they would have resembled the blow-wells along the Humbe Bank, where similar infiltrations of sea-water develops hydrostati pressures which have forced the underground Chalk-waters ninety fee upwards through Quaternary silts and clays to form blow-wells. Likewise it seems not unreasonable to assume that similar hydrostatic forces wer developed in the sands beneath the Herrin Coal to form the circula hollows around which the trees petrified and disintegrated into the surgin spring waters which would have been heavily charged with colloida exudates from the surrounding coal-forming swamp. Such exudates would not only lead to the precipitation of dopplerite but also enhance th solubilities of mineral matter.

The occurrence of marine deposits above the Herrin Coal is associate with deep channels at this and other horizons in the Pennsylvanian through out Southern Illinois (Wanless, 1955, and Potter & Siever, 1955). Takin into account the progressive contraction of the humus deposits it would only require a lowering of about six feet in land-levels to bring this marin transgression. Later work will probably show the existence of tidal inlet near these coal-ball localities which would have intensified the infiltratio of saline waters into the underlying aquifer of sands. Such conditions would be ideal for blow-well formation and the salt-marsh conditions for th localised petrification of the trees. It is not suggested that all coal-bal were formed in this way, but this example emphasises the probability tha coal-ball material was petrified by a process distinct from coal formatio

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A Fauna from the Puttenham Beds (Lower Greensand) of Hampshire

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BSTRACT: The Puttenham Beds of the north-west Weald are shown by their newly scovered fauna to belong to the *cunningtoni* Sub-zone of the *nutfieldensis* Zone (Upper ptian). The fauna is listed. The deposits of the *cunningtoni* Sub-zone in west Sussex ould also be called the Puttenham Beds. Deposition took place during a period of arine transgression.

1. THE PUTTENHAM BEDS

HE NAME 'Puttenham Beds' was proposed by Kirkaldy (1932) for the ivision of the Lower Greensand in the north-west Weald called the oamy Folkestone Beds' by the Geological Survey (Dines & Edmunds, 29). They have been described lithologically in the Guildford and odalming area by Dines & Edmunds (1929), in the Hascombe area by irkaldy (1932) and in the Hindhead area by Knowles & Middlemiss 59) but their precise stratigraphical position has been unknown in the sence of fossils.

The zonal position within the Aptian of the underlying Bargate Beds is xed by the rare occurrence of *Parahoplites nutfieldensis* Sowerby in the odalming area (Dines & Edmunds, 1929), at Green Cross near Hindhead ichardson, 1947) and near Midhurst, Sussex (Kirkaldy & Wooldridge, 38). Farther afield *P. nutfieldensis* occurs in Group XIII of the Ferrunous Sands (Lower Greensand) of Shanklin and Blackgang, Isle of ight. The species serves throughout this area to define a horizon equivant to the Nutfield Fuller's Earth Beds of East Surrey, the Sandgate Beds Kent and the Faringdon Sponge Gravels.

Overlying the horizon of *Parahoplites nutfieldensis* both in the Isle of ight and the western Weald is a horizon of loamy and ferruginous sands -Group XIV of the Isle of Wight, the so-called 'Sandgate Beds' of west ssex and the Puttenham Beds of Hampshire and west Surrey. The first o of these have for a long time past yielded fossils but the last have been regarded as unfossiliferous. Casey (1951) has shown that the overlyin Folkestone Beds in the north-west Weald contain ammonites indicative o the *Leymeriella tardefurcata* zone of the Albian. Thus it is of interest t decide whether the Puttenham Beds are of Upper Aptian or Lower Albia age and whether they can be correlated with the somewhat similar bed lying above the *P. nutfieldensis* horizon farther to the south.

2. THEIR FAUNA

The fauna listed below is the first to be described from the Puttenha Beds of the north-west Weald. It is entirely in the form of moulds and cast and comes from ferruginous sandstone nodules set in ill-graded sands no far above the base of the beds at Headleywood Farm (814372), Headley Hampshire (Knowles & Middlemiss, 1959). The locality was visited by th Geologists' Association in May 1957 (Middlemiss, 1959).

PORIFERA
- Spicules of undetermined genera and species

ECHINOIDEA
- *Phyllobrissus fittoni* (Wright)
- *Catopygus* cf. *columbarius* (Lamarck)
- *Holaster* sp. (cf. *benstedi* Forbes sp. or *wrighti* Lambert)

POLYZOA
- Polyzoan of undetermined genus and species

VERMES
- *Serpula* sp.

BRACHIOPODA
- '*Rhynchonella*' sp.

LAMELLIBRANCHIA
- *Limatula tombeckiana* (d'Orbigny)
- *Thetironia minor* (J. de C. Sowerby)
- *Freiastarte* cf. *subcostata* (d'Orbigny)
- *Resatrix* (*Resatrix*) *parva* (J. de C. Sowerby)
- *Venilicardia* sp.
- Cf. *Nemocardium* (*Pratulum*) *ibbetsoni* (Forbes)
- *Protocardia* cf. *sphaeroidea* (Forbes)
- *Protocardia* sp.
- *Modiolus aequalis* (J. de C. Sowerby)
- *Gervillella* sp.
- *Pterotrigonia* cf. *mantelli* Casey
- *Parmicorbula striatula* (J. de C. Sowerby)
- indet. corbulid
- *Exogyra tuberculifera* Koch & Dunker
- *Ostrea?*
- *Toucasia lonsdalii* (J. de C. Sowerby)
- *Gyropleura* sp.

GASTROPODA
- *Margarites* (*Atira*) sp.
- *Dimorphosoma* sp.

MONOIDEA
Parahoplites cunningtoni Casey
Parahoplites sp. juv. cf. *melchioris* Anthula
Parahoplites sp. juv. cf. *schmidti* Sinzow non Jacob
CERTA SEDIS
Granularia sp. (faecal pellets)

The presence in numbers of *Limatula tombeckiana* and *Margarites* *tira*) sp. is a local character. The latter is closely allied to undescribed ecies in Group IV of the Ferruginous Sands at Atherfield, Isle of Wight, d in the Folkestone Beds. *Gyropleura* is very rare in the Lower Greensand t does occur in the Folkestone Beds at Farnham. The occurrence of the dist *Toucasia lonsdalii* is noteworthy as this is the only rudist known in e Lower Greensand and has been found only at this horizon (Casey, 61, 578). The fossils of greatest stratigraphical value are the ammonites hich, apart from some very rare *Parahoplites* in the ironstone nodules of roup XIV of the Ferruginous Sands at Shanklin, Isle of Wight, are the ly ammonites so far known from this horizon in the Lower Greensand. he specimens have body chambers and are therefore not the nuclei of rger examples but represent juvenile casualties. Their small size prevents rect comparison with the Group XIV specimens, but there is little doubt at they are part of the same fauna.

3. CONCLUSIONS

The ammonites indicate the *Parahoplites cunningtoni* Sub-zone of the *rahoplites nutfieldensis* Zone (Casey, 1961), at the top of the Upper ptian.

The remainder of the fauna, although exhibiting some local features, is osely comparable with that of the 'Sandgate Beds' of Pulborough, Sussex irkaldy, 1937), and of Group XIV at Shanklin and Blackgang in the Isle Wight (Fitton, 1847; Reid & Strahan, 1889). Thus throughout the area om the south-west of the Isle of Wight to the north-west of the Weald the ds lying above the horizon of *P. nutfieldensis* and forming the uppermost vision of the Aptian are alike in containing ferruginous nodules with this una. Kirkaldy's (1939) belief that these outcrops are all of the same rizon has been confirmed.

It has always been customary to refer to this horizon in west Sussex as e 'Sandgate Beds'. However, in the type area of east Kent this name operly belongs to all the beds between the Hythe Beds below and the lkestone Beds above, that is including all the beds characterised by *rahoplites*. In west Surrey and Hampshire the Sandgate Beds can be early subdivided into the Bargate Beds with *Parahoplites nutfieldensis* low and the Puttenham Beds with *P. cunningtoni* above. In west Sussex

P. nutfieldensis is known from the Bargate Beds, thus it is illogical to use the term 'Sandgate Beds' for the overlying *P. cunningtoni* horizon only. Consequently it is here proposed that the nomenclature employed in the north-west Weald should be applied to west Sussex—the name Puttenham Beds to be used for the deposits probably referable to the *P. cunningtoni* sub-Zone, including the Selham and Pulborough Iron Sands, the 'Sandgate Loams', with their included fossiliferous ironstones, and the Pulborough Sand Rock. The name Sandgate Beds should be retained to cover both Bargate Beds and Puttenham Beds.

The Puttenham Beds fauna is of notably different facies from that of the Bargate Beds. In the Puttenham Beds lamellibranchs and gastropods are the commonest fossils, while brachiopods are rare and terebratulid brachiopods absent. Ammonites, although not yet found at this horizon in Sussex, are yielded by many of the fossiliferous nodules in Hampshire and the Isle of Wight. In the Bargate Beds, on the other hand, terebratulid brachiopods and echinoids are the dominant fossils, while lamellibranchs are represented mainly by oysters. Ammonites, although represented in the Bargate Beds by four species from a variety of localities, are not common fossils in these beds, since they occur in very much smaller numbers per unit volume of fossiliferous sediment than in the Puttenham Beds of Hampshire and the equivalent Beds of the Isle of Wight. This difference is correlated with the marked lithological differences between the divisions—the Bargate Beds are generally coarse, calcareous, moderately glauconitic, strongly current-bedded; the Puttenham Beds are loamy, ill-graded, ferruginous, almost invariably decalcified, strongly glauconitic, only locally current-bedded. The Puttenham Beds do not exhibit any facies indicative of littoral or sub-littoral conditions of deposition anywhere in the Weald, whereas the Bargate Beds pass, between Godalming and Guildford, into a coarsely bioclastic pebbly facies showing every indication of approach to shore. The Puttenham Beds therefore were probably deposited during a phase of the Upper Aptian transgression on to the London landmass and represent conditions farther from shore than the Bargate Beds.

ACKNOWLEDGMENT

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The ammonites have been presented to H.M. Geological Survey. The remainder of the fossils are in the author's possession at Queen Mary College, University of London.

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A Rapid Method for Making Plaster Casts for Geological Work

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T IS OFTEN necessary to reproduce large numbers of geological specimens plaster of Paris for use as teaching aids and examination material when ere is a shortage of natural specimens.

Numerous excellent methods for mould making have been described but nfortunately these methods are time-consuming, an important consideraon in a busy teaching department where a small technical staff has to deal ith a great variety of work. For this reason it was found necessary to evise a rapid and simple technique which would give accurate reproducon of detail and, when possible, dispense altogether with multi-mould ethods and their complicated keying and supporting procedures.

One of the difficulties encountered in moulding natural materials is the ct that they are often porous. Wood, for example, contains large quanties of air and water, while fossils preserved in loosely compacted sediientary rocks are extremely porous and therefore contain large amounts f air.

When making a mould it is usual in the process to heat the original ecimen, either by baking it in the moulding material or by pouring the quefied material over it. In either case the air contained in the specimen cpands and bubbles up through the moulding material leaving the inner ırface of the mould pitted with small cavities on cooling. These obliterate ıuch of the surface detail required on the plaster casts. Normally this ifficulty is overcome by treating the specimen with several coats of sealing ompounds, with long drying periods between each application. Quite part from the fact that this is a lengthy process, many geologists object to ieir specimens being treated in this way.

The partial-vacuum technique evolved by the writer has dispensed with ıe necessity for sealing porous material, and only large internal cavities, ıch as those in gastropods, need be plugged with Plasticine.

Using the modified methods to be described, complicated wooden ıodels of crystals, and such difficult fossils as *Hemicidaris*, *Siphonia* and *entriculites*, have all been faithfully and quite rapidly reproduced in laster of Paris. Up to thirty accurate casts free from flash lines have been btained from one mould.

In all cases the moulding material employed has been 'Vinamould' Grade HMT 1028), a flexible rubber-like material which, when melted, is poured round the specimen and allowed to solidify. 'Vinamould' is capable of reproducing the finest detail on plaster casts. It can be obtained, together with a special heating vessel and an informative booklet on moulding techniques, from Messrs. Vinatex Ltd., Carshalton, Surrey.

The 'Vinamould' is prepared for use by cutting it into small cubes which are then placed in the heating vessel to melt. The heating time depends on the amount of material required; a full container of five pints capacity will require heating for about three hours before the contents are ready for pouring.

The writer employs two methods in making a mould, depending on the texture of the specimen to be reproduced.

Method 1 (for non-porous specimens)

This is relatively simple and is used for non-porous specimens which have either at least one flat surface as, for example, a crystal of gypsum, or an unimportant surface such as the underside of the rock matrix adhering to a fossil.

The specimen is placed on a smooth glass or metal surface with the important surface uppermost. If the specimen has an uneven under-surface it may be levelled by pressing it into a bed of Plasticine, care being taken that the latter does not protrude beyond the area covered by the base of the specimen.

A strip of Plasticine is then placed round the specimen to form a ring leaving a gap of two or two and a half inches between the Plasticine and the specimen. Next, a simple dam of strong but flexible cardboard is superimposed on the Plasticine wall; office file covers will be found suitable for this purpose. The card is rolled to form a tube and the overlapping edges sealed together, inside and out, with a gummed adhesive strip. The diameter of the dam is determined by that of the Plasticine ring, and the height is governed by that of the specimen; it is necessary for the dam to be at least two inches higher than the specimen. Fig. 1 A illustrates the method of setting up.

The mould is now ready to receive the melted 'Vinamould'. This is poured in until the top of the specimen is covered with about one and a half inch of 'Vinamould'. The latter should not be poured over the specimen itself, but directed into the space between the specimen and the dam in a steady, uninterrupted flow. Afterwards a few gentle taps on the side of the dam will dislodge any trapped air which will then rise to the surface. The completed mould should now be set aside to cool, preferably overnight.

When the material has solidified the dam and its contents are pulled away from the base plate, and the cardboard and surplus Plasticine removed from the mould. The specimen is now gently eased out of the mould which will be found to be elastic and capable of being stretched to facilitate oval of the enclosed specimen. At this stage the interior of the mould uld be examined for imperfections. If it is satisfactory the first of the ster casts can be made from it.

The above method for making moulds may also be employed for non-rous, irregularly shaped specimens where *all* the surfaces are important;

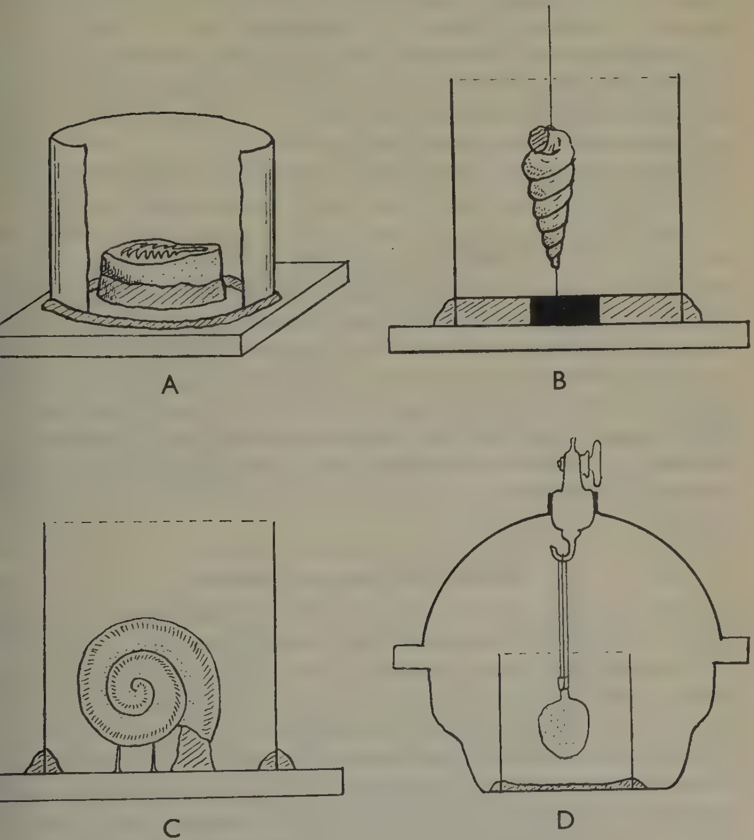

Fig. 1. Setting-up specimens for moulding

A diagrammatic representation of a specimen supported on Plasticine with the dboard dam in position

Gastropod suspended by a light thread and anchored to weight embedded in the sticine

Ammonite supported on pins with aperture filled with Plasticine extended to form ther support

Siphonia Tulipa suspended in desiccator by thread and held in position by glass ing enclosing thread to prevent specimen from rising but in such cases a certain amount of ingenuity may be required in setting up the specimen. The supporting material must not leave an impression in the mould thereby interfering with the detail reproduced on the casts. The difficulty is usually overcome by supporting the specimen on pins (long map-pins are useful for the purpose) or by suspending the specimen on fine cotton from a suitable support. Fig. 1 B illustrates methods used for supporting specimens. When the 'Vinamould' has set, the dam and supporting structures are removed, and the mould carefully cut open with a sharp knife to reveal the specimen inside. The split must be just big enough to facilitate removal of the specimen. The interior of the mould should be inspected for flaws before use.

When making plaster casts in such moulds, the split must be held open while the plaster of Paris is poured in, care being taken to ensure that the two sides of the split go back into accurate register when closed after pouring. As an added precaution, the sides of the split should be dusted with french chalk, and a strong elastic band placed round the mould to ensure that correct register is maintained while the plaster of Paris hardens.

It sometimes happens that a small unimportant portion of the specimen is present which may be utilised to help support it and to provide a pouring vent. For example, where the aperture of an ammonite is filled with consolidated matrix, the latter may be extended outwards in Plasticine as a supporting structure. When the cast of the specimen is made this extension is also reproduced, but it is easily broken off without affecting the appearance of the rest of the cast.

Method 2 (for porous specimens)

The second method is employed for porous specimens, and it is extremely useful for reproducing wooden models of crystals.

It is generally similar to Method 1, but differs in that it employs a vacuum desiccator and a simple water-type vacuum pump. A twelve-inch desiccator has been found useful for most purposes.

The bottom of the desiccator is lined with a layer of Plasticine, and the specimen placed in position, the method of support being the same as in Method 1 for those specimens where a cast of the entire surface is required. In the case of crystal models made in wood, however, there is a tendency for them to float to the top of the hot 'Vinamould'. This can be prevented by pressing one of the flat surfaces of the model into the Plasticine so that it remains firmly held down when the 'Vinamould' is poured in. Light, porous fossils also have a tendency to rise to the surface but they may be satisfactorily anchored as illustrated in Fig. 1 D.

The cardboard dam can now be placed in position and pressed firmly into the Plasticine lining the desiccator. The hot 'Vinamould' is poured in to cover the specimen to a depth of about one and a half inch. The lid of the siccator (slightly greased to ensure a good vacuum seal) is replaced and e stop-cock checked to make sure that it is in the 'off' position.

The vacuum pump is then connected and the stop-cock slowly opened itil the molten 'Vinamould' is seen to rise and bubble gently; this manipu- tion must be carefully controlled so that the 'Vinamould' is allowed to se until it is about half an inch from the top of the dam. The stop-cock is en closed and the vacuum maintained for three or four minutes. After is the vacuum may be slowly reduced until the 'Vinamould' sinks back ain. The lid of the desiccator is removed and the contents topped with ore liquid 'Vinamould' to within half an inch of the top of the dam.

The procedure is now complete. As before, the mould is left to cool, and finally made ready for use by removing the specimen as described in ethod 1.

Preparation and Handling of Plaster of Paris

Plaster of Paris is the material normally used for the preparation of sts; it is simple to mix and accurately reproduces fine detail. Colouring aterial may be introduced to the mix, or the hardened cast may be rface painted on completion.

Numerous trade brands of fine plaster are available on the market; they nd to be relatively expensive, and little advantage is gained by buying ese brands for ordinary purposes. 'Industrial fine' plaster of Paris gives rfect cast detail and can be purchased for less than twenty shillings per ndredweight.

Plaster hardeners may be added when mixing; these reduce the setting ne and produce extremely hard and durable casts. 'Vinatex IS/6' is a eful hardener for this purpose.

To ensure perfect casts it is suggested that the following procedure be opted. The moulds are first washed under running water to remove dust d loose material from the inner surface, and the plaster of Paris prepared. common fault is to mix excessive amounts; to avoid this the moulds ould be filled to capacity with clean water which is then poured into a bber mixing-bowl, add to this the plaster of Paris by sprinkling small antities on to the surface of the water. Water added to plaster will pro- ce a lumpy mixture. Continue adding the plaster until a crust is formed; this stage the mixture is smoothly stirred until it has the consistency of eam. This mixture will flow easily during pouring, and will run into all e grooves and undercuts of the detail in the mould.

After thickening, the plaster of Paris is poured into the mould which ust rest on a level surface. When filled to the top the plaster is poured ck into the basin and then repoured into the mould. The first pouring is rticularly important when small moulds are used, since it ensures a ating of plaster on all the undercut and cavity surfaces in the mould. In both cases the pouring should be made slowly and carefully, the rubbe basin being squeezed to facilitate pouring.

After filling the moulds they should be tapped to encourage air to rise t the surface. Afterwards open moulds and pouring vents should be seale by placing over them pieces of clean glass. Excess plaster of Paris must b wiped off before the casts are allowed to harden.

ACKNOWLEDGMENT

The author extends his thanks to Mr. J. D. R. Fryer for reading thi paper and making helpful comments.

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WRITTEN CONTRIBUTIONS TO THE DISCUSSION OF PAPERS PREVIOUSLY TAKEN AS READ

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e Editor

ear Sir,

r. H. G. Owen is to be congratulated on his work (*Proc. Geol. Ass., Lond.*, **71**, 4) describing six sections related to the Gault–Lower Greensand junctions in e Eastern Section of the Maidstone By-Pass (1960), and on their stratigraphic rrelation which confirms Dr. R. Casey's (1958, 1959) conclusion that the Lower ault sequence in the district is complete.

The writer was concerned in the area during the re-survey which preceded the blication in 1958 of the New Series One-Inch Geological Map of the Maidstone strict. In the course of this work the strike-fault which was recognised as fecting the Gault–Folkestone boundary between Aylesford and Pennenden eath, appeared locally to be reversed and locally to pass into a monocline.

Confirmation of the reversed nature of the fault was obtained in January 1958 ring the early stages of the by-pass construction, when five boreholes were nk by the Kent County Council's Roads' Department Laboratory at sites jacent to the fly-over which now carries the by-pass over the Chatham Road rid ref. TQ 754585). The first borehole was drilled on the western verge of the atham Road and the remainder on the eastern side. The cores obtained from e boreholes and the pits in which they were sunk were examined conjointly by e staff of the Roads' Department Laboratory and the writer. The summaries of e sections and their classification, given below, are by the writer.

rehole No. 1 (105 yards south of the entrance to Zoo Park)

	Thickness ft.	*Depth ft.*
Made Ground	4	4
Folkestone Beds: Brown sand with irregularly cemented, calcareous sandstone at top	16	20
Brown sandy clay with streaks of blue-black clay (Fault breccia)	2	22
Gault Clay: Blue-grey clay with gypsum, sandy and with phosphatic nodules at base	14½	36½
Folkestone Beds: Brown-yellow sand ...	1+	37½

rehole No. 2 (45 yards east of Borehole No. 1)

	Thickness ft.	*Depth ft.*
Soil	¾	¾
Folkestone Beds: Brown, clayey sand, grey-green at base	14¼	15
Hardened light grey clay (Fault breccia) ...	5	20
Gault Clay: Blue-black clay, sandy and green at base	18	38
Folkestone Beds: Green-yellow sand ...	1+	39

Borehole No. 3 (40 yards south, 60° east of Borehole No. 1)

	Thickness ft.	*Depth ft.*
Made Ground	3	3
Folkestone Beds: Yellow-brown sand with irregularly cemented, calcareous sandstone at base	1+	4

Borehole No. 4 (50 yards, south 70° east of Borehole No. 1)

	Thickness ft.	*Depth ft.*
Made Ground	4	4
Folkestone Beds: Yellow-brown sand with irregularly cemented, calcareous sandstone in top 9 ft.	10¾+	14¾

Borehole No. 5 (60 yards, south 80° east of Borehole No. 1)

	Thickness ft.	*Depth ft.*
Soil	½	½
Folkestone Beds: Brown sand with irregularly cemented, calcareous sandstone at base	7½+	8

The successions in Boreholes 1 and 2 demonstrate that they passed through reversed fault which results in Folkestone Beds at the surface overlying Gaul Clay, resting in normal succession on the Folkestone Beds below.

From fossils collected from Borehole 2, Dr. Casey identified *Falcifere milbournei* Casey (25 ft. and 25 ft. 1 in.) indicative of the *Anahoplites intermediu* Subzone of the *Hoplites dentatus* Zone, together with *Hoplites* aff. *dentatus* (J Sow.) and *H.* aff. *spathi* Breistroffer from 32 ft.; the latter indicative of the *H* *dentatus-spathi* Subzone of the *H. dentatus* Zone. This small part of the succession is normal, and being situated only 400 yards west of Owen's (1960) Section can be taken as further evidence that the Lower Gault succession in the area i complete.

When the fly-over was subsequently constructed a cutting was made whic passed through the sites of the boreholes. Although the exposures were poor the writer was able to amplify the evidence provided by the trial boreholes and to observe a low-angled thrust (3° to 5°), which at the southern end of the cuttin steepened rapidly to hade southwards at approximately 30°.

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THOR'S REPLY

e author would like to thank Mr. Gray for communicating the information tained from the exploratory boreholes sunk in the vicinity of the A.229 atham road at the east end of the West Section of the Maidstone By-Pass and result of his work on the cutting. The reversed nature of the fault which he has monstrated at this locality becomes less marked eastwards, but is still apparent the Grange Lane Cutting Section 5 (1960, 372). The disturbance of the Gault, hough marked at the faulted southern edge of the outcrop, does not extend to y great distance northwards. Although the exposure of the Lower Gault was t extensive in Section 6 at the Boarley Bridge (1960, 372), it showed little dis- bance and possessed a dip of about 2° northwards. Farther to the south-east the north-west quadrant of the A.249 clover-leaf Section 4 (1960, 369), the gle of dip was found to be but a few degrees towards the NNE. near the junc- n of the quadrant with the A.249 Sittingbourne road. It increased, however, 25°–30° near the edge of the Gault outcrop which is cut by three minor strike lts with throws in the direction of dip. The southern fault affects the Folke- ne Beds and the Gault–Lower Greensand Junction only while the other two ect the Lower Gault. On the north side of the By-Pass 200 feet west of the nction of the north-east quadrant and about 700 feet along the strike from ction 4, a single strike fault with a dip throw of a few inches could be seen and s affected the Gault–Lower Greensand Junction only. On the south side of the -Pass this fault was no longer apparent (1960, 366). The officers of the Geo- gical Survey, who at that time did not have the benefit of the By-Pass sections, curately terminated the strike fault at the point represented on Sheet 288. The ting of the By-Pass at this locality has shown, however, that the southern edge the Gault outcrop does not swing eastwards in the manner they indicated but ntinues farther on a strike bearing of 126° (true) before turning eastwards.

The principal fault which affects the southern edge of the Gault outcrop, refore, becomes more important as one proceeds north-westwards from the 249 Clover Leaf towards Aylesford.

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e Editor

ar Sir,

a recent paper, 'Pleistocene Sea-levels in North Devon' (*Proc. Geol. Ass.*, *nd.*, **71**, 169), the author, Miss M. A. Arber, states that two low-level surfaces erosion are present, the Hele Surface (150 feet plus to about 100 feet plus) and Croyde Surface (at about 50 feet). Both these surfaces are attributed to marine nation of a periglacial deposit, the head, and the high sea-levels are stated as, t likely to be of a later date than the last Interglacial, and the head may there- e belong to the Penultimate Glaciation. The head itself is younger than the cial erratics of North Devon, which may therefore belong to a still earlier ciation' (p. 175).

There can be no doubt whatever that the head deposits (the plural is used iberately) are younger than the large glacial erratics of North Devon, because can be shown that the head deposits overlie both the large erratics and the sed beach deposits in which many of the erratics lie. G. F. Mitchell has recently olished a paper (*Advanc. Sci.*, 1960) in which it is suggested that the Main Lower Head dates from the Riss or Saale glacial period (the Penultimate aciation of Zeuner, 1959).

Recent investigations show that the Main Head of the coastal seotions around Barnstaple Bay is a frost rubble deposit of great thickness, representing the product of a period of severe periglacial climate. The material consists of blocks of all sizes of slate, sandstone and quartz, set in a sandy matrix, and where it has not been subsequently disturbed by later frost action the rock fragments have a preferred orientation downslope, often lying at gentle angles to the horizontal and projecting out of the face of the cliff when seen in section. From the higher ground inland the marerial has moved down the old convex coastal slope as a kind of sludge, and spread out as a great 'apron' or solifluxion terrace with a concave inner margin at the foot of the old coastal slope. The surface of this 'apron' tends to lie at very low angles of less than three degrees of slope, forming an almost level 'shelf' or terrace. This is well seen at Saunton Down and Baggy Point (south side). The surface of this terrace is always composed of head, sometimes overlying considerable thicknesses of raised beach shingle and sandrock (well-bedded sands, with shingle lenses and layers). No marine deposits have been found to cap the Main Head, and in no place can the coastal 'aprons' or solifluxion terraces be shown to owe their final 'terrace-like' form to marine planation.

This is just as true in the river valleys draining to the coast, where the rivers have trenched the head deposits which flowed off the valley's sides to fill the valleys to a considerable depth; very careful analysis of the sections is necessary to distinguish between the true river (alluvial) terraces and the dissected solifluxion deposits, especially when the aim is to make interpretations as to the levels attained by sea-levels during the Pleistocene.

Furthermore, on the coast around Barnstaple Bay, at Saunton Downend and Middleborough, for example, the surface of the Main or Lower Head has been disturbed by later frost action; in places it is overlain by a layer of sand, above which another layer of Younger or Upper Head can be seen; and the Upper Head may consist of multiple layers of frost rubble; frost wedges can be seen to dissect the upper head at a number of points near Middleborough. Capping all these head deposits is a variable thickness of sand and slate fragments, perhaps with some small pebbles, this representing a wash of material from higher up the slope, and this uppermost layer is known to contain Mesolithic flint flakes.

Close examination of these surfaces thus shows them to be without marine deposits, and their morphology to be very uneven, with frequent changes of altitude; and at many places the heads can be traced in unbroken sequence from the present cliff section up slope across the surfaces to the old coastal slope at the rear of the solifluxion terrace. If further proof is needed then it may be pointed out that similar terraces can be observed in many parts of southern England, south Wales and southern Ireland, that is, in areas which lay, for the most part, outside the limit of the glacial advances of the Riss (Saale) and Würm (Weichsel) periods; the phenomena has been recorded by many authors.

In point of detail, the 'Hele Surface' 'around Hele and Bickington' as reported by Miss Arber (p. 171) is not cut across head deposits; the Hele–Ellerslie–Bickington rock ridge is capped by sands and gravels, in places probably some 30–40 feet thick at least; the deposits show strong cryoturbation, convolutions and frost wedges, but are in primary position. Mitchell (1960) has suggested that they represent an aggradation phase of the Taw–Torridge system in the early Pleistocene, and the frost action probably took place during the Riss (Saale) glacial period.

The Anchor Wood river cliff (p. 172) cut in rock on the left bank of the Taw downstream of Barnstaple, cannot have any significance in separating the Hele and Croyde surfaces; it is simply the undercut edge of the Hele–Bickington rock ridge.

The Croyde Surface (p. 172) is essentially a solifluxion terrace in the Croyde lley; and around Fremington (lying to the south of the Hele–Bickington ge) the land between 50 and 200 feet consists for the most part of solifluxion posits—sandy, stony sludges, not unlike some boulder clays—which at the emington clay pits can be seen to cap the chocolate boulder clay. Only at the ast near Fremington Quay can the raised beach gravels be seen to reach the rface, but here there is no typical head of slate rubble—the ground surface nply consists of the exposed raised beach (its surface deeply disturbed by oturbation), which can be seen in section to the west of Fremington Quay in e cliffs, in the railway cutting to the east, and in the cliff near the old lime kiln Penhill Point. This raised beach, equivalent to the beach sealed below Main ad on the outer coast, can be traced southwards below the Fremington ulder clay near Combrew Farm. Thus the 50 foot contour lies partly upon sed beach material (which is older than the boulder clay, and the Main Head the outer coast) and partly upon solifluxion material which also seals off the e Fremington boulder clay. There is no trace whatever of an old cliff-line cut her in the solifluxion materials or in the boulder clay near Fremington, yet a nce at the Ordnance Survey map will show that this area would have been mpletely open to a marine incursion which exceeded 50 feet O.D.

To sum up, there is, in my opinion, no evidence for the planation of the head marine transgressions, and the sequence of events around Barnstaple Bay may her be stated as follows:

st-Glacial	Alluvium, hillwash; Mesolithic implements.
ürm/Weichsel	Younger Upper Head and frost wedges.
erglacial	No trace of marine deposits.
ss/Saale	Disturbance of surface of Main Head (frost cryoturbation).
	Main Head.
	Fremington Boulder Clay.
erglacial	Raised beach (deposit from 20–60 feet O.D.).
ndel/Elster	Large erratics reach the coast (Saunton Red Granite; Croyde (Freshwater Gut) granite-gneiss).
rly Pleistocene	Cutting of the low rock platforms below 50 feet on which the large erratics rest.
	Hele–Ellerslie–Bickington sands and gravels.

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THOR'S REPLY

nade clear in my paper on 'Pleistocene Sea-levels in North Devon' that I ered my own conclusions simply as working hypotheses until more evidence s forthcoming, and I welcome the fact that Mr. Stephens and Mr. Mitchell ve already carried so much further the study of what I termed the Hele and oyde Surfaces, as well as having investigated the stratigraphical details of the head deposits. The major difference between Mr. Stephens's interpretation an mine of the sequence of events lies in his belief that the Hele and Croyde Surface are subaerial aggradation platforms, whereas, in spite of the absence of stratigraphical evidence, I thought them to be more probably of marine origin.

In North Devon, aggradation terraces are particularly well seen south of Hartland Point. At Marsland Mouth, the double terracing (at approximately 100 fee and at 50 feet) indicates two stages of rejuvenation post-dating the head deposits when each successive valley-floor was formed, sea-level must have stood highe than at present, and I hope that Mr. Stephens will publish evidence of the heigh and of the position of the coast-line at these stages.

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